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PETROGRAPHY AND ORIGIN OF THE TUSCARORA, ROSE HILL, AND KEEFER FORMATIONS, LOWER AND MIDDLE SILURIAN OF EASTERN WEST VIRGINIA[1]

ROBERT L. FOLK
The University of Texas, Austin, Texas

ABSTRACT

The Tuscarora Formation is a medium sandstone with a few layers of vein quartz pebbles. The lower, red portion is a subgraywacke to graywacke, and was deposited in a submerged deltaic-estuarine environment. The upper, white portion is a supermature orthoquartzite formed as a beach sand deposited during a period of declining relief. The source area lay to the southeast; during early Tuscarora deposition it consisted largely of older sediments and low-rank metamorphic rocks, but as erosion proceeded, the sedimentary cover was stripped off and the plutonic igneous sources became of increasing importance. The Rose Hill Formation consists largely of open-marine to brackish-water olive clay shale, but intercalated with this are thin layers of gray siltstone and red hematitic sandstone, ranging from orthoquartzite to subgraywacke in composition. The hematitic sandstones were deposited in a more shallow marine oxidizing environment than the gray siltstones which are cemented with ankerite and quartz. Iron minerals are common in the Rose Hill, and hematite, chlorite pellets, ankerite, and pyrite show consistent environmental associations. The source area for the Rose Hill consisted of low-rank metamorphic plus plutonic igneous rocks. The Keefer Sandstone shows evidence of lagoon-border and beach-dune deposition and consists of immature to supermature orthoquartzite to subgraywacke sandstones. The source area had now become deeply eroded so that plutonic igneous rocks dominated, but there were still some low-rank metamorphic rocks exposed.

High rounding was accomplished in one cycle of deposition in the beach and dune environments of the upper Tuscarora and upper Keefer sandstones; rounding potency varied markedly from bed to bed. Metamorphic rock fragments, because of their low resistance to abrasion, are found to be very useful as environmental indicators.

INTRODUCTION

Much of the petrographic stratigraphy of the north-central Appalachians has been outlined under the guidance of P. D. Krynine at Pennsylvania State University. Units studied petrographically (chiefly in central Pennsylvania) include: Upper Cambrian Gatesburg sandy dolomite (Pelto, 1942); Lower Ordovician Beekmantown carbonates (Folk, 1950, 1952; Rones, 1956); Bellefonte Sandstone (Krynine and Tuttle, 1941b); Middle Ordovician limestones (Folk, 1946); Upper Ordovician and Lower Silurian Oswego, Juniata, and Tuscarora Sandstones (Tuttle, 1940; Krynine and Tuttle, 1941a); Lower Devonian Oriskany sandstones (Rosenfeld, 1953); and Middle Devonian Chemung sediments (Krynine, 1940; Haney, 1952). Much of this material was summarized by Krynine (1946) who integrated the history of changing source areas and tectonic events.

This long section, almost completely covered by petrographic studies, contains one conspicuous undescribed gap: the Middle and Upper Silurian. This part of the section has remained unstudied because it has very little sandstone and consists chiefly of mudrock and carbonate. But many intriguing rock types are present: pea-green, massive mudstone (Wills Creek), two different types of redbeds (Rose Hill and Bloomsburg), and gray or black shale (Rochester, Wills Creek) along with more ordinary shale, limestone, and thin sandstone beds. The writer became greatly interested in deciphering source and depositional environment of this part of the Silurian through contact with the stratigraphic research of F. M. Swartz and during one summer's field work in the Silurian of central Pennsylvania under C. A. Bonine. However, the Silurian section in the eastern panhandle of West Virginia was chosen for the present study because of the many good outcrops and the very detailed stratigraphic work done there already by Woodward (1941); furthermore, comparison could be

[1] Manuscript received June 11, 1959.

made between results here and those obtained by previous studies under Krynine in central Pennsylvania.

The Silurian of northeastern West Virginia is a generally transgressive geosynclinal sequence, starting with thick conglomerates and sandstones resulting from the Taconic Orogeny in southern New England, progressing through shales and thin sandstones as this highland was worn down, and grading into carbonate sections as peneplanation and extensive marine transgression occurred. One renewed pulse of terrigenous material (Bloomsburg redbeds) interrupted this transgression. Quiescent conditions continued well into the Devonian with deposition of limestone and thin orthoquartzite blankets (Oriskany; Rosenfeld, 1953) before the immense wave of graywacke sandstone, shale, and redbeds caused by the Acadian orogeny in the upper Devonian (Chemung and Catskill). Thus the rocks studied occupy that part of the section between the climax of the Taconic Orogency and the midpoint of the long quiescent interval preceding the Acadian Orogeny (Swartz, 1948).

This study was undertaken to decipher the changes that took place in source area paleogeology between the sedimentary and low rank metamorphic source of the upper Ordovician (Tuttle, 1940; Krynine and Tuttle, 1941a; Krynine, 1946) and the primary plutonic source of the lower Devonian (Rosenfeld, 1953). A second purpose was of more general petrologic interest. Almost no information exists on the properties of shales as studied in thin section, and seldom has the petrography of shales and limestones been integrated with that of the associated sandstones in an attempt to determine source area, environment of deposition, or paleoclimate. The properties of the thin limestones that occur in geosynclinal sequences are all but unknown. The writer feels that the story is incomplete without studying all these varied rock types, and therefore decided to examine what is essentially one vertical section of the entire Silurian System to trace the sequence of changes throughout the entire geosynclinal transgression. This first publication covers the lower, almost entirely terrigenous part of the system, including the Tuscarora sandstone and conglomeratic sandstone, Rose Hill shale and siltstone, and Keefer sandstone. The next paper will describe the Rochester black shale and the McKenzie shale and limestone; and the final paper will cover the Bloomsburg redbeds, Wills Creek claystone and limestone, and Tonoloway limestone and dolomite.

All sections were studied in the vicinity of the eastern panhandle of West Virginia in Berkeley and Morgan Counties or in outcrops along the Potomac on the Maryland side (Washington County). Single complete sections of each formation were measured except for the McKenzie in which two sections had to the spliced together; sections were located from the descriptions of Woodward (1941). Because only one section of each formation was measured, this report does not consider lateral facies changes nor persistence of the formational subdivisions; it is intended to set up a "type petrographic section" of the Silurian System for this region in the hope that other type petrographic sections will be set up elsewhere.

METHODS

Sections were measured in the field and proportions of various lithologies (for example, siltstone or limestone beds in shale sequences) were estimated for each interval. In sandstone or siltstone sections, grain size was determined by the use of a comparison set of sand grains spaced at 0.5ϕ intervals. With this device it is possible to estimate median grain size in the field to within $\pm 0.15\phi$ of its value as obtained in the laboratory. Mineral percentages and gravel content were obtained in the field by a percentage comparison chart (Folk, 1951a). In the laboratory, detailed descriptions were made of each hand specimen with binocular microscope; the National Research Council Rock Color Chart was used, interpolating when necessary. Limestone specimens were described both on fresh breaks and on acid-etched surfaces. Insoluble residues were obtained quantitatively and studied by binocular and petrographic microscopes. Representative specimens of each lithic type were selected for thin-sectioning, and if a given lithic type was often repeated in the section, thin-sections of it were made at many stratigraphic levels.

The terminology used in describing sandstone and shale specimens follows the grain size, textural maturity, and mineral composition terminology of Folk (1951b, 1954, 1956), while that for limestones follows his carbonate terminology (Folk, 1959a).

Percentages of major constituents were obtained by 100-point count of thin sections. A different method was used for constituents less than approximately 5 percent. A microscope ocular was used with a grid containing 10×10 squares. In each field of view, one can count how many squares or fractions thereof are occupied by the grains of each minor constituent, counting only those within the area of the 10×10 grid, and this becomes a direct areal measure of percentage. These results were averaged for ten fields of view.

In studying the Silurian, the eight formations were described petrographically in random, not stratigraphic, order to insure against any bias inasmuch as long-range vertical changes in mineralogy and texture were to be studied. Furthermore, within each formation the thin-sections were done in random instead of stratigraphic order for the same reason. Considering things as subjective as estimating mica content in shales of the Rose Hill, for example, if all the shale slides had been done in stratigraphic order the measured increase in mica upward might have been caused by a subconscious operator drift in definition of mica, but because the slides were done in random order, the consistent high mica content of the Upper Rose Hill can be confidently ascribed to a real change and not to operator error. Similarly, the upward decrease in metaquartzite and chert in the Tuscarora is much more certain because of the random order in which the slides were described.

TUSCARORA SANDSTONE

Regional Relations

Rocks of the Lower Silurian, designated as the Albian or "Medinan" Series, represent the culmination of the Taconic Orogeny if one considers the great areal extent of coarse gravel deposits and the enormous mass of sandstone produced during this epoch. In the northern Appalachians, no volume of sandstone anywhere near approaching this was to be laid down until late Devonian, and no other important gravel beds were to occur until the Mississippian. The Lower Silurian is represented throughout this area by the Tuscarora Sandstone (or "quartzite"), the most spectacular ridge-former in the Ridge and Valley Province of the Appalachians. Forming the escarpment that borders the Great Valley and the Shenandoah Valley on the west almost all the way from New York to Tennessee, it also holds up most of the prominent mountain ranges to the west as far as the edge of the Allegheny Plateau. This thick and very hard, white to light gray sandstone is made up of almost pure quartz sand and gravel cemented very tightly with silica so that bare outcrops form many scenic crags, and great quantities of float blocks mantle the mountainsides. Figure 1 shows the regional variation of this mass of sediment; in northern New Jersey the Tuscarora is equivalent to the lower part of the Green Pond Conglomerate, a pinkish, unfossiliferous milky-quartz pebble conglomerate at least 1500 ft thick, possibly of continental origin, lying unconformably on Cambro-Ordovician limestone and even Precambrian gneiss (Kümmel, 1940). In eastern Pennsylvania and southern New York it corresponds to the lower part of the Shawangunk (Swartz and Swartz, 1930, 1931), which consists of up to 1900 ft of sandstone and conglomerate with abundant white quartz pebbles. Passing southwest the Shawangunk grades insensibly into the Tuscarora, a very hard, pure and uniform white sandstone with scattered zones of white quartz pebbles, especially in the lower part; it varies from massively bedded to cross-bedded. The Tuscarora maintains this character from Pennsylvania as far as southwest Virginia, and there are pebbly zones even as far south as Bluefield, West Virginia, and Roanoke, Virginia (Miller and Fuller, 1954). Amsden (1955) presents an isopach and lithofacies map of the Lower Silurian and shows that in general *Arthrophycus* borings occur throughout the sandy facies (Tuscarora), while eurypterids occur in black shale within the conglomerate facies (Shawangunk) although in eastern Pennsylvania both fossils occur in a transition zone (Swartz and Swartz, 1930). Eurypterids are interpreted as fresh-water

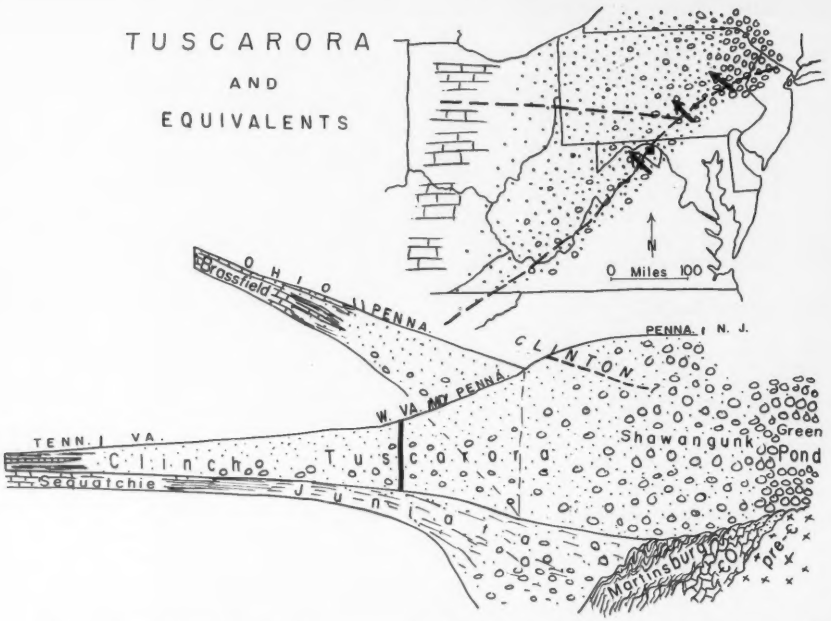

FIG. 1.—Regional variation in the Tuscarora Sandstone and equivalents. The formation is highly conglomeratic in eastern Pennsylvania and New Jersey and contains some pebbles as far as southwestern Virginia. The sandstone grades into limestone and shale to the west and southwest. Location of the section studied for this report is shown by the black square in the eastern West Virginia panhandle and by the heavy vertical line in the fence diagram; it is approximately 300 ft thick and has a few pebbly zones. Arrows show generalized direction of transport based on the crossbedding studies of Yeakel (1958). These appear to indicate that the Tuscarora was not derived from a point source in the vicinity of southern New York but rather fron a linear source that extended in a south-southwesterly direction and hence gradually diverged from the present eastern limit of the outcrop.

arthropods; consequently, Swartz (1948) considered the Shawangunk to be partly non-marine in the east.

In southern Virginia the Tuscarora is termed the Clinch Sandstone, and in far southwestern Virginia the first important interbeds of shale and fossiliferous limestone appear. The Clinch Formation extends into southeast Tennessee before it disappears by faulting, and the equivalent beds in northern Alabama are calcareous and fossiliferous sandstone, some limestone, shale, and minor hematite ore beds.

Going west from Pennsylvania and West Virginia, the sands of the Tuscarora thin and finally disappear in eastern Ohio (Rittenhouse, 1949; Amsden, 1955), and the equivalent in western Ohio and Kentucky is considered as the fossiliferous Brassfield Limestone which grades into dolomite in Indiana and Michigan (Amsden, 1955).

The isopach pattern (Woodward, 1941) shows an axis of maximum thickness trending from central Pennsylvania through extreme western Maryland into central West Virginia. The maximum thickness along this axis ranges from about 700 ft in central Pennsylvania to 300–400 ft in western Maryland and northern West Virginia to 100 ft in southern West Virginia and southwest Virginia. The measured thickness varies erratically in western Maryland, northeastern West Virginia, and northern Virginia, ranging from less then 50 ft (or even less than 5 ft, Butts and Edmondson, 1939) to almost 400 ft even in this small

area (Butts, 1940). These erratic thicknesses depend partly upon the difficulty of choosing a consistent boundary between it and the underlying red Juniata Graywacke. Swartz (1934) shows that the Lower Tuscarora has some pink beds well up into the formation that are similar to the Juniata; the same difficulty occurs in West Virginia where hard Tuscarora-looking sandstone beds and reddish Juniata-appearing beds are intercalated over a sometimes broad transition zone (Woodward, 1941). Butts and Edmondson (1939) thought the variable thickness might be caused by irregular deposition of the Tuscarora, but Giles (1942) considered that it was caused by obscure strike faulting that cut out parts of the sections.

Thus the formation shows a rather simple picture: a great mass of conglomerate in New Jersey and eastern Pennsylvania, and the massive, moderately thick pebbly sandstones in central Pennsylvania, West Virginia and Virginia grading into thinner and more limy marine phases south and west. This conspicuous wedge of terrigenous sediment is attributed to the Taconic Orogeny which became very active in late Ordovician and was probably more pronounced in southern New England or southeastern New York (Kay, 1942), but a series of smaller uplifts probably took place at several points farther south along the borderland generally termed Appalachia. In other words, the formation was probably not derived from one "point" source, but more likely from a "line" source that was perhaps highest in the north. Crossbed measurements by Yeakel (1958) show that a northwesterly transport direction prevailed for the Tuscarora throughout the area from northern West Virginia to eastern Pennsylvania, and this favors the idea of a linear source extending parallel with the present outcrop belt.

Underlying the Tuscarora at most places is the red Juniata graywacke which is thickest and coarsest with some pebbly beds in central Pennsylvania (Swartz, 1948). Generally it is a red sandstone and shale (for example, Queenston Delta), but it grades into marine Richmond shales and limestones in western New York, Ohio, and east Tennessee (Sequatchie).

From Harrisburg, Pennsylvania, southwest there is no disconformity beneath the Tuscarora (Woodward, 1941); but at Swatara Gap northeast of Harrisburg the Tuscarora unconformably overlies the Martinsburg Slate and the Juniata is not present. Pebbles of slate are incorporated into the base of the Tuscarora, and evidence shows that the slaty cleavage was imposed prior to erosion of the pebbles (Woodward, 1941). The time interval represented by the unconformity increases eastward until the Silurian lies on Precambrian gneiss.

The Tuscarora section studied for this paper is about 40 miles southeast of the axis of greatest thickness at the easternmost outcrop belt for this latitude. Measuring along strike, it is about 150 miles southwest from beds that are dominantly conglomerate and 250 miles northeast of the first limy and fossiliferous beds.

Local Character

Field relations. The Tuscarora Sandstone is excellently displayed along a roadcut on West Virginia Highway 51 immediately south of the intersection of this road with West Virginia Route 45 at Mills Gap on Little North Mountain; the section is located in Gerrardstown Township, Berkeley County, West Virginia, about 10 miles southwest of Martinsburg. As the road climbs the moutainside it exposes a nearly continuous section from the Ordovician Martinsburg Shale and Juniata Formation through the Tuscarora Sandstone and into the lower Rose Hill Shale. The formations are here overturned, the beds dipping approximately 45° to 50° southeast and striking N 40° E; the attitude is nearly constant with no minor folds and no evident faults of more than a few inches throw. Samples were taken at 5-ft intervals except in concealed zones.

As is common in other places, the difficulty in this section lies in choosing the contact between the Tuscarora Sandstone and the underlying Juniata redbeds (Ordovician), a series of graywacke sandstones and shales. Three sharply defined lithologic zones are present. Zone 1, at the base, consists of only moderately hard, non-resistant, usually thin-bedded muddy subgraywacke sandstone and abundant fissile mud-shale,

which are red both on fresh and weathered surfaces. Zone 1 passes abruptly into zone 2 which is approximately 65 ft thick, composed of very hard, highly-resistant, quartz-cemented medium- to thick-bedded subgraywacke sandstone beds (with no shale), which are red on a fresh break but normally weather with a whitish rind; consequently the outcrop looks like typical Tuscarora. This zone contains many red clay galls and three levels showing *Arthrophycus* imprints. Above this zone is a concealed interval of approximately 30 ft, and the next exposures belong to zone 3, approximately 215 ft thick, which consists of light gray to white, medium- to thick-bedded, very hard, quartz-cemented, clean sandstone and gravelly sandstone mostly of orthoquartzite composition. Zone 1 is unquestionably Juniata and zone 3 is unquestionably Tuscarora, but zone 2 could conceivably be assigned to either formation. If it is considered as Tuscarora, the 65 ft of sandstone in zone 2 and the 30 ft of concealed interval overlying it would make the Tuscarora 310 ft thick, much thicker than the formation as measured by Woodward (1941, p. 61), who obtained only 130 ft for the Tuscarora on the northeast side of Mills Gap (but 100 ft of section was covered above his described Tuscarora). Massive bedding characteristics, lack of intercalated shales, extreme hardness, whitish weathered color and general appearance of the outcrop ally it with the Tuscarora; yet it is similar to the Juniata in its red color on a fresh break, usual presence of a muddy matrix (which the Tuscarora lacks), and the subgraywacke mineralogy of the detrital grains. Thus zone 2 represents a transition between the two formations. Nevertheless, because of the presence of *Arthrophycus* (usually thought to be a diagnostic fossil for the Tuscarora) and because of the general field appearance, this writer chooses arbitrarily to place zone 2 in the Tuscarora although the evidence is about equally weighted on both sides. Thus the base of the Tuscarora is chosen at the first appearance of massively bedded, very hard and resistant, white-weathering sandstones above a section of red-weathering, softer sandstones and intercalated red shales; this contact is 0.24 mile south of the concrete marker at the intersection of West Virginia Highways 51 and 45.

The top of the Tuscarora is considered as the topmost outcropping ledge of quartzite; above this a zone of nearly 100 ft of float before the first Rose Hill shales are observed in place.

The basal 65 ft of the formation (zone 2 as discussed above, hereinafter called the "red Tuscarora") is followed by about 30 ft of concealed interval. This concealed zone is believed to consist of thin-bedded reddish platy sandstones on the basis of examination of an equivalent part of the section also poorly exposed on the roadcut going northeast from Mills Gap. Thus the red Tuscarora would total approximately 95 ft. Some rather poor exposures of white, gravelly Tuscarora succeed the concealed zone and continue for about 25 ft stratigraphically (that is, between 95 and 120 ft above the formation base). These outcrops are followed by another concealed zone approximately 45 ft thick; at 165 ft above the formation base good outcrops of white sandstone reappear and the section is continuous from there to 310 ft.

Because of the radical textural and mineralogic differences between the lower, red Tuscarora and the upper, white Tuscarora, they will be discussed separately.

Lower (Red) Tuscarora

The lower part of the Tuscarora Formation at Mills Gap consists of a series of beds that on a fresh break have a color between 5R4/1 and 5R4/2 (grayish red; color designations after the National Research Council Color Chart, Goddard and others, 1948); a few beds are lighter when fresh, approximately 10YR7/1 (yellowish gray). The grayish red sandstones usually develop an outer weathered zone 1 to 15 mm thick that is bleached much lighter to about 5R8/2 (grayish pink). Thus, these rocks have very nearly the same color on the weathered surface as the white, upper Tuscarora. The red color is due to hematite, and it has been bleached out since the roadcut was made.

The beds of the red Tuscarora are hard to very hard (about the same as the white Tuscarora) but differ somewhat in type of bedding. The bulk of this unit occurs as

sandstone beds 3 to 6 in thick, but interbedded with these is about 5 to 10 percent of beds of softer, very platy sandstone that splits easily into slabs $\frac{1}{4}$- to $\frac{1}{2}$-in thick. Both rock types are essentially similar in color and grain size, but the softer thin-bedded sandstones tend to be more subject to cross-bedding. The bulk of the 30-ft concealed interval above the known red Tuscarora is thought to consist of this platy sandstone.

All beds in the lower, red Tuscarora are thinly laminated (1–2 mm) with layers of slightly differing sand grain size, and laminae are straight and smooth with no evidence of curdling or slumping. About half of the beds show well-developed cross-bedding with amplitudes of 3 to 6 in; once the tectonic folding is compensated for, the cross-beds dip quite consistently in a west to northwest direction, indicating currents flowing from the southeast.

Grayish red clay galls (5R4/2) are very abundant in the red Tuscarora, usually concentrated in sandstone beds an inch or so thick. These disc-shaped, well-rounded clay pebbles lie parallel with the bedding, and most are 1 or 2 in in diameter and about $\frac{1}{4}$-in thick, but some are as large as 5 in. The galls occur selectively along the partings which define the individual sandstone beds; thus large bedding surfaces are often exposed on which one may see these galls or the pits from which they have been removed. Clay galls disappear abruptly on passing into the upper, white Tuscarora. *Arthrophycus* impressions were found at three levels within the red Tuscarora.

Microscopic characteristics.—The average red Tuscarora sample is described as a muddy medium sandstone: siliceous, immature subgraywacke (Folk, 1954, 1959b). Texturally, the red Tuscarora varies between medium and fine sandstone, and about half of the specimens have a muddy matrix (plate 1, A, B) while the others are relatively "clean" (plate 1, C). No gravel is present at all; in fact the size of the largest grains observed in individual thin sections normally ranges between 0.8 and 1.2 mm (see Textural Variation Chart, fig. 2). The modal grain size of the sand varies between .18 and .35 mm (1.5 to 2.5ϕ), typically being .25–.30 mm. Sand grains are moderately to well sorted and apparently unimodal even up to and including the coarsest quartz particles. The sand grains are mixed with 5 to 20 percent of silt and clay matrix. Sands range between immature and submature[2] depending on clay content; clay content and sorting may vary widely between laminae even within the same slide.

Quartz grains average subangular but vary widely between angular and subround; the roundest grains occur at a size of .20–.25 mm (Plate 1, D). A few grains that were once almost perfectly rounded have been split in half during transportation (plate 1, E). All slides examined show the same mixed assemblage of roundness, and because of the abundant presence of angular grains, apparently all the observed roundness has been inherited from past environments or older formations. The high roundness at .20–.25 mm indicates that this must have been the modal grain size of one of the sources, perhaps an older supermature quartz sandstone. Most of the silt grains (.03 to .06 mm) are also rather well rounded. It is interesting that the metamorphic quartz grains tend to be more angular than the grains of common (non-strained) quartz which have been largely inherited from older sandstones.

Compositionally, rocks sampled are on the borderline between subgraywacke and graywacke[3] (plate 1, B, D), ranging between 15 and 35 percent M pole constituents and averaging 23 percent. The main mineral variety is common quartz[4] which forms be-

[2] Textural maturity scheme is as follows (Folk, 1951b, 1954, 1956): sands with more than 5 percent clay matrix are classed as immature; those with smaller amounts of clay but poor sorting are submature. Sands with good sorting (σ under 0.5ϕ) are mature, and well-sorted and rounded sands are supermature.

[3] Graywacke is defined as a terrigenous rock whose gravel-sand-silt grains consist of more than 25 percent M pole constituents (metamorphic rock fragments, discrete micas, and stretched composite metaquartzite) and less than 10 percent feldspar. This definition is independent of clay content, sorting, cementation, induration, geologic age, color, or mode of formation. Subgraywacke contains 5 to 25 percent of these constituents (Folk, 1954, 1956.)

[4] Common quartz (Folk, 1959b): single crystallographic units with straight to slightly undulose extinction and few vacuoles; the "plutonic quartz" of Krynine (1940).

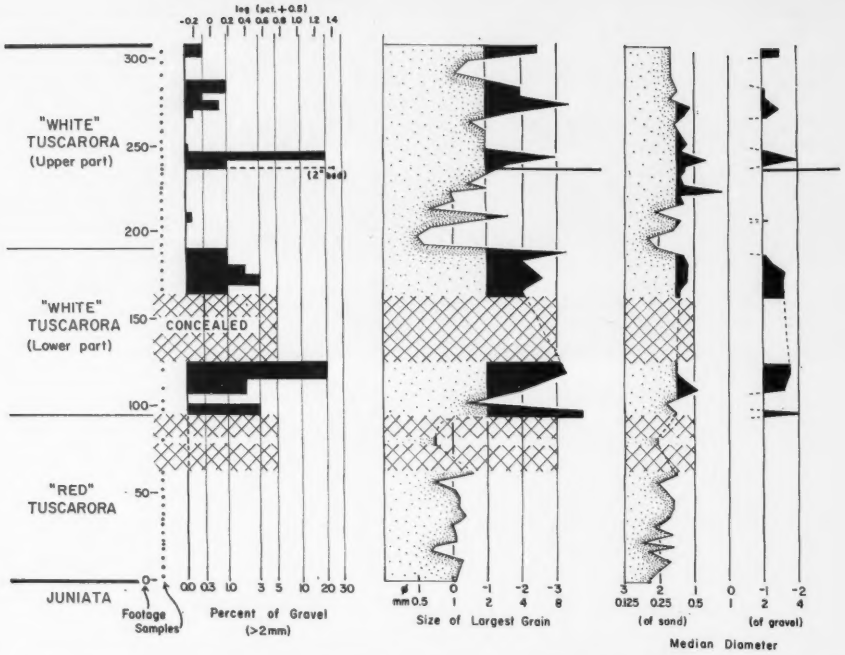

FIG. 2.—Textural variation in the Tuscarora Formation at Mills Gap, Berkeley County, West Virginia. Concealed zones shown by cross-hatching. The first column shows the percentage of gravel on a modified logarithmic scale. There is no gravel in the red Tuscarora but considerable amounts in the lower part of the white Tuscarora and some in the upper white Tuscarora. The second column shows the size of the largest grain encountered in each thin section or hand specimen. On this basis the red Tuscarora has the smallest maximum grain size (mostly about 1 mm), and the lower white Tuscarora has the coarsest grains. The third column shows the median diameter of the specimens or, in gravelly samples, the median of the gravel fraction and the median of the sand fraction estimated separately. Medians in the red Tuscarora are about 2ϕ, those in the white Tuscarora about 1 to 1.5ϕ.

»»→

EXPLANATION OF PLATE 1

A. Sandstone with hematitic clay matrix, texturally immature. A dark clay lamina crosses the bottom of the picture. Note prevalent angularity of quartz and occasional well-rounded quartz grains inherited from older sediments. A few slate and metaquartzite fragments are visible. The rock is a typical subgraywacke, almost a graywacke. Crossed nicols, ×25.

B. Immature graywacke with hematitic clay matrix and poorly sorted, angular grains. Metamorphic rock fragments (speckled grains) are abundant. Crossed nicols, ×32.

C. Submature subgraywacke. Hematitic clay matrix has been winnowed out but the grains are still poorly sorted and angular. Pore space is completely filled with quartz overgrowths. Crossed nicols, ×25.

D. Typical graywacke sandstone of the red Tuscarora, textually immature because of a red clay matrix. Note common small rounded quartz grains derived from older sediments; large grains are angular. Metamorphic rock fragments are abundant. Crossed nicols, ×25.

E. Immature graywacke with a large broken round grain derived from older sediments. Crossed nicols, ×25.

F. Immature graywacke showing abundant low-rank metamorphic rock fragments, probably slates. Note their elongate shape caused by foliation or cleavage in the parent metamorphic terrane. Crossed nicols, ×86.

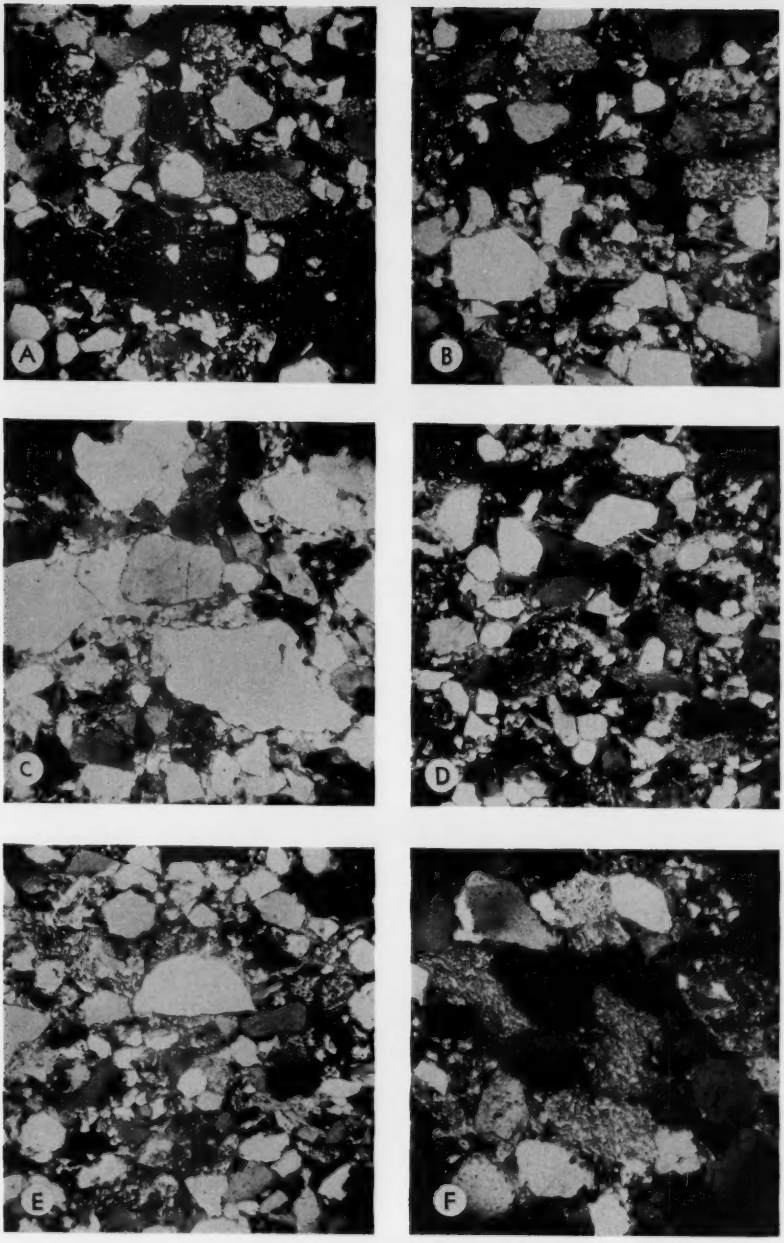

PLATE 1.—Photomicrographs of lower (red) Tuscarora Sandstone.

tween 50 and 70 percent of the terrigenous fraction. About half of these grains show undulose extinction; many of them are strongly undulose. About 7 to 15 percent of straight-extinction, composite quartz is present, most of it made up of a fine-grained (20 μ) quartz mosaic and probably derived from recrystallized metamorphic sources. There is 3 to 6 percent chert in subround to subangular fragments about the same size as the associated quartz; some of the chert contains abundant illite inclusions. In all the thin sections only one grain of feldspar (a sericitized orthoclase) was seen.

The most diagnostic rock constituents are the soft metamorphic-rock fragments (plate 1, B, F), which comprise 12 to 25 percent of the sandstone samples. These consist very largely of weakly-metamorphosed clay-slates composed of 5–10 μ sericite with minor chlorite. They show good foliation and mild recrystallization, but were still soft enough to have been bent around neighboring quartz grains upon compaction. The grains are elongated and of about the same size range as the associated quartz. Although the clay matrix of these sandstones and the clay galls are thoroughly permeated with hematite, the metamorphic rock fragments have usually escaped this taint except for surficial staining. The sandstones contain 4 to 10 percent stretched metaquartzite in addition to the slate fragments. These grains are made up of several elongate, lenticular quartz crystal units showing undulose extinction and usually having crenulated internal boundaries. It is surprising that despite all these evidences of a low-rank metamorphic source there are no discrete mica flakes (that is, flakes longer than about 20 μ). In addition to these definite metamorphic rock fragments, the unit also contains 1 to 5 percent of orthoquartzite-siltstone fragments which are very weakly if at all metamorphosed. These fragments are somewhat larger than the single quartz grains (about .30–.50 mm) and are made up of well-sorted subangular to subround .03 mm quartz silt cemented with hematite and quartz.

Heavy minerals are chiefly subround magnetite with some leucoxene; these two minerals sometimes occur as intergrowths in the same grain. Tourmaline and zircon are very scarce. The only difference between the red and the white Tuscarora in heavy mineralogy is the presence of magnetite in the redbeds and its absence in the white unit. This situation also is true if one compares the red and the gray siltstones and sandstones in the overlying Rose Hill; Miller and Folk (1955) explained this as having been caused by post-depositional solution of magnetite from the more reducing units while it was preserved in the oxidized (red) units.

The contained clay galls consist of hematite-stained illitic clay together with some quartz silt. The clay is well-oriented but shows no evidence of any recrystallization or metamorphism; together with the large size of the galls, this indicates that they are simply bits of red, oxidized flood-plain or tidal-flat mud that have been torn up and redeposited shortly after they were first laid down. They are easily distinguished from the detrital slate fragments.

Matrix, when present, consists of illitic clay and some silt rather heavily impregnated with hematite (plate 1, A). In the laminae that are washed clean of clay, quartz cement occurs as overgrowths and fills the available pore space almost completely, at places amounting to 5 percent of the rock (plate 1, C).

The hematite in the red Tuscarora has formed as a chemical precipitate after deposition of the sand which was previously unstained. If the red color was due to the reworking of a red, deeply weathered soil in the source area, one would have expected the metamorphic rock fragments to be heavily weathered and permeated with hematite; instead they are free of any internal stain and only the margins are coated with iron oxide. In many places quartz grains that touch lack hematite at the points of contact, indicating that the hematite came in after deposition of the unstained sand grains. Formation of very thin hematite rims around all grains was followed by development of quartz overgrowths as the final phase of diagenesis. The hematite probably originated by weathering of iron-bearing minerals (such as biotite, amphiboles, and pyroxenes) in the metamorphic and igneous rocks of the source terrane.

Upper (White) Tuscarora

The white part of the Tuscarora Formation is approximately 215 ft thick, extending from 95 to 310 ft above the formation base. The bedding is somewhat more massive than that of the lower, red Tuscarora, averaging 6 in to 1 ft thick quite consistently although some beds as thin as 3 in or as thick as two ft are encountered. The rocks are hard to very hard, and indeed the field term "sedimentary quartzite" could be applied inasmuch as most of them are so tightly cemented with silica that they have little or no porosity and break through the grains. The color is quite constant; approximately half the specimens lie between 10YR8/1 and 10YR8/3 (very pale orange) because of faint limonitic stain; the other half are N8/5 to N9 (white). Specimens become slightly friable upon weathering and develop a dark yellowish orange to light brown (5YR to 10YR6-7/6) limonite-stained crust approximately $\frac{1}{2}$-in thick. The cause of this stain is not known since there is no apparent source of this much iron within the rock; perhaps it may be infiltrated from surrounding soil.

The upper 20 ft of the formation contains a few scattered green clay galls, and occasionally in the rest of the formation sandstone beds will be separated by a thin parting of greenish clay, not more than a mm or so thick. Other than these very minor occurrences, detrital clay is essentially lacking in the white Tuscarora.

The beds generally maintain constant thickness along the outcrop face, and in nearly all of the sections prominent 1–3 mm lamination is expressed within these units. These laminae are near-horizontal and are caused by slight changes in sand grain size. In most of the section, cross-bedding can be faintly seen in approximately one bed in four. It is usually of small amplitude and rather gentle dip; however, between 260 and 280 ft, cross-bedding is very well expressed in sets ranging from 1- to 12-in thick. Throughout the section there is a weak but general tendency for most of the cross-beds to dip between west and north (in agreement with the results of Yeakel, 1958), although if all the beds are considered, the dips virtually box the compass. Even in one of the most strikingly cross-bedded units at 268 ft, beds dip in all directions. This much more random direction of cross-bedding is in sharp contrast with the consistent west to northwest dip of cross-beds in the red Tuscarora below.

Texturally, the upper Tuscarora is the coarsest and best winnowed formation in the entire Silurian section studied (fig. 2). Some gravel (grains coarser than 2 mm) is found in more than half of the samples, although in most of these it makes up no more than a trace to 2 or 3 percent of the hand specimen. The gravel is estimated to make up 1.5 to 2 percent of the entire section sampled. These pebbles and granules are embedded in a matrix of medium sand; the rocks are cleanly washed with clay and even silt lacking in almost every specimen. The gravel occurs in two zones in the Mills Gap section. The most important gravelly zone begins at the base of the white Tuscarora (95 ft above the formation base) and extends to 190 ft. Here the gravel is most abundant and also coarsest. A second gravelly zone occurs at 235 to 285 ft. Pebbles consist largely of well-rounded, milky to clear vein quartz, although subangular chert pebbles are common toward the base of the section.

In the lowest gravel zone, virtually all the outcropping beds contain some gravel over a stratigraphic thickness of nearly 100 ft, although most of the middle of this section is concealed. Most specimens have only 1 to 5 percent gravel,[5] but one sample attains 20 percent. The median of the gravel fraction averages 3 to 4 mm, and the largest pebble observed in each bed generally ranges from 4 to 10 mm (fig. 2). Most specimens are classed as slightly granular medium sandstone, although several beds are

[5] Gravel percentages were obtained in the field by using a percentage comparison chart (Folk, 1951a); sand grain sizes were also estimated in the field with a set of mounted standard grains spaced at 0.5ϕ interval. After thin sections had been described, sizes obtained in thin section were compared with those determined in the field, and a small correction factor (-0.1 or -0.2ϕ) was added to the field designations to bring them into agreement with the thin section data which was assumed to be more correct. For example, if the field estimate of the median was 1.4ϕ, the "true" size is 1.2ϕ or 1.3ϕ; these corrected values are the figures used in the text.

just over the border into pebbly medium sandstone; grain size of the sand matrix is 1.2–1.5ϕ (.35–.45 mm).

The lower gravel zone is succeeded by a zone of pure sandstone 45 ft thick (except for one bed having 0.1 percent of granules). Other beds contain no gravel and are mostly well-sorted medium sandstones averaging very close to 1.4ϕ (.40 mm); but there are a few beds of fine sandstone (2.3ϕ, .21 mm) and one bed of very well-sorted coarse sandstone (median 0.3ϕ, 0.80 mm). Maximum size grains observed in samples of this zone range from 1ϕ (0.5 mm) at the base to -1ϕ (2 mm) at the top.

The second gravel zone is 50 ft thick; some thin beds in this interval contain as much as 20 percent gravel but most have only a trace to 1 percent and are classed as slightly granular medium sandstone. In the middle third of this interval the maximum grain size hardly if at all attains 2 mm. The gravel median is mostly 2–3 mm with the largest pebbles being 4 and 10 mm, but one bed at 235 ft (the lowest sample in this gravel zone) has a median pebble size of 9 mm with the largest pebble 18 mm. In all specimens the granules and pebbles are embedded in a matrix of medium sandstone (1.2–1.5ϕ, 0.35–0.45 mm).

This is succeeded by the final thin zone (285–305 ft) of well sorted medium sandstones (1.7ϕ, .32 mm) with maximum grains averaging –0.5ϕ (1.5 mm).

Texturally then, the upper (white) Tuscarora could be divided into two cycles; each cycle showing a gradation from gravel upward into sand, with the gravel being most coarse and most abundant at or near the base of each gravelly zone. This could be considered as very large-scale graded bedding, indicating a sudden influx of gravel followed by a gradual decline. The upper gravel zone is less gravelly and contains smaller pebbles than the one below.

Microscopic characteristics.—In the lower part of the white Tuscarora, the average sample is a slightly pebbly or granular medium sandstone (siliceous submature chert-bearing quartzose subgraywacke), while in the upper part it is a medium sandstone (siliceous supermature orthoquartzite). The median grain size of samples averages 0.37 mm or 1.45ϕ, with a tendency for gravelly samples to have slightly coarser sand although the difference is slight (0.2–0.5ϕ). In the pure sandstones, two-thirds of the grains in any slide usually range between .20 and .50 mm; consequently, samples are on the borderline between well-sorted and moderately sorted (plate 2, A). The smallest grains observed in the slides (except for wisps of authigenic clay) range from .05 to .08 mm; thus, silt and detrital clay are almost totally lacking in the upper Tuscarora.

Most of the samples are bimodal and many are polymodal. The following modes

EXPLANATION OF PLATE 2

A. A sandstone in the lower part of the white Tuscarora showing typical degree of sorting (good, but not extreme). A few composite grains (meta-quartzite) are visible. The rock is tightly cemented with quartz overgrowths. It is supermature (barely so) and borderline between an orthoquartzite and a quartzose subgraywacke. Crossed nicols, ×25.

B. Typical medium sandstone of the upper part of the white Tuscarora, an extreme example of a supermature orthoquartzite with excellent sorting and rounding. It consists almost entirely of common quartz, tightly cemented by quartz overgrowths so that grain roundness is invisible in this photograph. Crossed nicols, ×8.5.

C. Usual poor sorting of one of the gravelly sand beds in the lower part of the white Tuscarora. Sediment is polymodal and each mode is vaguely defined. Grains are not very well rounded. This rock is classed as a submature orthoquartzite. Crossed nicols, ×8.5.

D. Submature (not well-sor ed) sandstone in the lower part of the white Tuscarora. The specimen is classed as a quartzose subgraywacke because of the abundance of metaquartzite (composite grains); a large elongate grain of chert is also present. Mineralogy is representative of this part of the Tuscarora. Crossed nicols, ×8.5.

E. Texturally mature (well-sorted) quartzose subgraywacke common in the lower part of the white Tuscarora. Three grains of stretched metaquartzite (arrows) are shown. Crossed nicols, ×25.

F. Chert granule near the base of the white Tuscarora. It contains abundant sericite flakes and some quartz silt and shows usual blocky shape with rounded corners. At this stratigraphic level, chert grains are common and much larger than associated quartz. Crossed nicols, ×25.

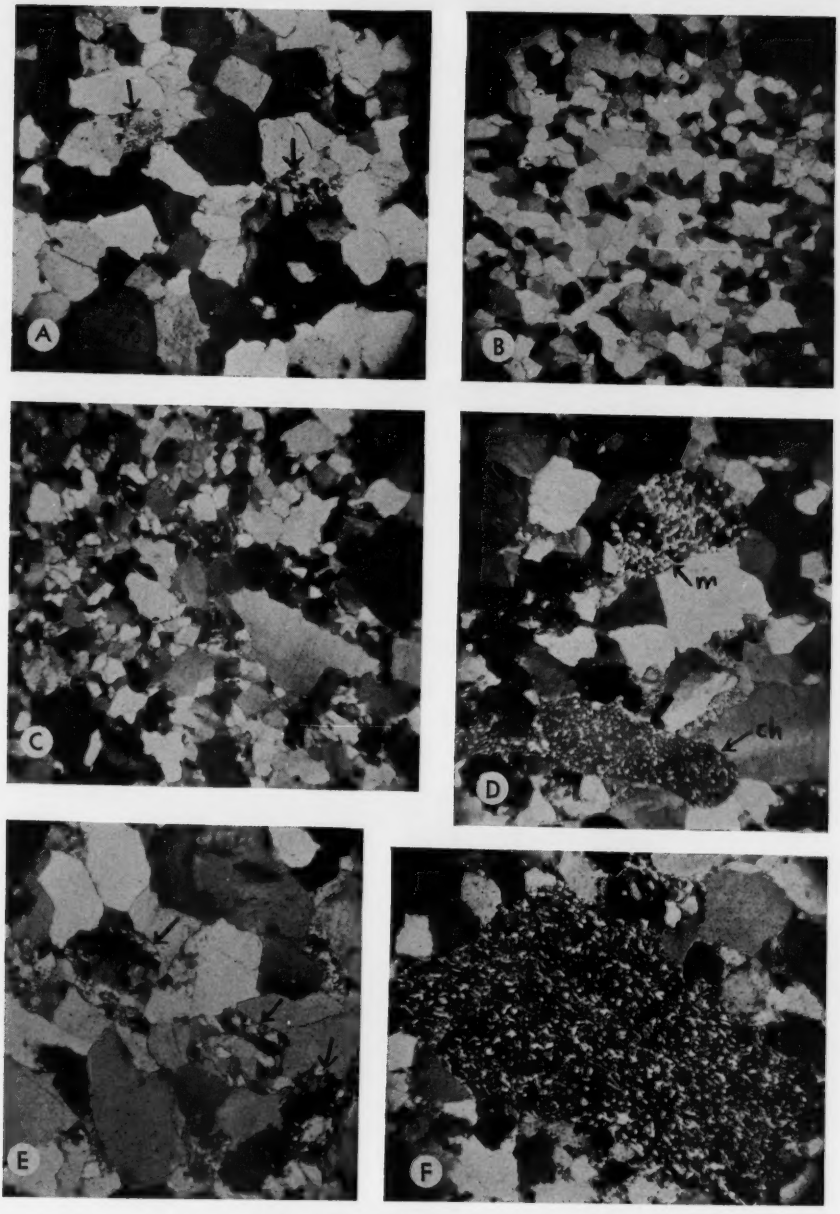

PLATE 2.—Photomicrographs of upper (white) Tuscarora.

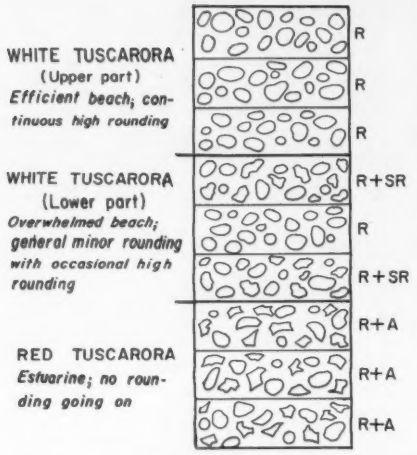

FIG. 3.—Diagrammatic representation of roundness of quartz in the Tuscarora Formation. In the red Tuscarora, all beds show the same mixed assemblage of roundness, with rounded grains inherited from older sediments plus angular grains from primary rocks. No rounding was being accomplished by the depositional agent as shown by the constant sharp angularity of the non-inherited grains. In the lower part of the white Tuscarora abrasion was sufficient to round all the grains to at least "subround" and at times produce beds in which all grains were well rounded. Thus the rounding potency of the environment was variable. In the upper white Tuscarora intensive beach action rounded all the grains thoroughly. This diagram intends to show that high roundness may be accomplished in one cycle of deposition by a potent environment.

seem to be most prevalent and distinct: .20–.25 mm (usually a minor mode), .40–.50 mm (usually the major mode), and 1–2 mm (occurring only in the gravelly samples). A few gravelly samples have additional modes at 5–10 mm. Most of the well-sorted samples are essentially unimodal medium sandstones (plate 2, B), and the more poorly sorted have three or more indistinct modes. The majority of the gravel beds are poorly sorted and polymodal, with the individual modes themselves commonly weakly defined and poorly sorted (plate 2, C). Usually the various modes are concentrated in certain layers (causing the laminated hand specimen appearance) and the pebbles, especially, accumulate in streaks parallel with the bedding.

From 90 through 190 ft (lower part of the white Tuscarora; that is, the entire first gravelly zone) the sand grains vary in mean between subround and round in different specimens; above 190 ft, grains are almost uniformly round to well-rounded (fig. 3).

With regard to textural maturity, all the sand-size grains are subround or rounder; thus the rocks would all be classed as supermature but for the fact that many of the beds are not well sorted. Consequently, the lower dominantly gravelly part of the section is classified as submature (inverted because of the rather high roundness), the middle portion shows an alternation of submature and supermature beds, and the upper portion consists largely of supermature sediments (plate 2, B) since it is largely pure well-sorted sandstone. This brings out the important point that the maturity concept as originally applied (Folk, 1951b) is without much meaning in conglomeratic rocks since nearly all conglomerates are bimodal with a sandy matrix and hence are automatically poorly sorted regardless of the amount of grain rounding.

Stratigraphic changes in mineralogy.—Mineralogically, the Tuscarora shows a marked vertical zonation in the Mills Gap section. As has been stated previously, the lower (red) Tuscarora is a subgraywacke to graywacke with abundant soft slate or phyllite fragments. The lower part of the white Tuscarora (from about 95 to about

EXPLANATION OF PLATE 3

A. Large chert pebble (ch), free of clay and silt, and broken by tectonic fracturing in the source area. Vein quartz pebble (v) showing typical semi-composite extinction. Crossed nicols, ×24.
B. Grain of recrystallized metaquartzite; note lack of stretching. Crossed nicols, ×24.
C. Grain of more coarsely crystalline recrystallized metaquartzite (m). Crossed nicols, ×24.
D. Grain of stretched metaquartzite; note elongate shape of individual crystal units and of entire fragment. Crossed nicols, ×24.
E. Pebble of vein quartz with typical semicomposite extinction. Crossed nicols, ×24.
F. Pebble of vein quartz with typical comb structure and semicomposite extinction. Crossed nicols, ×24.

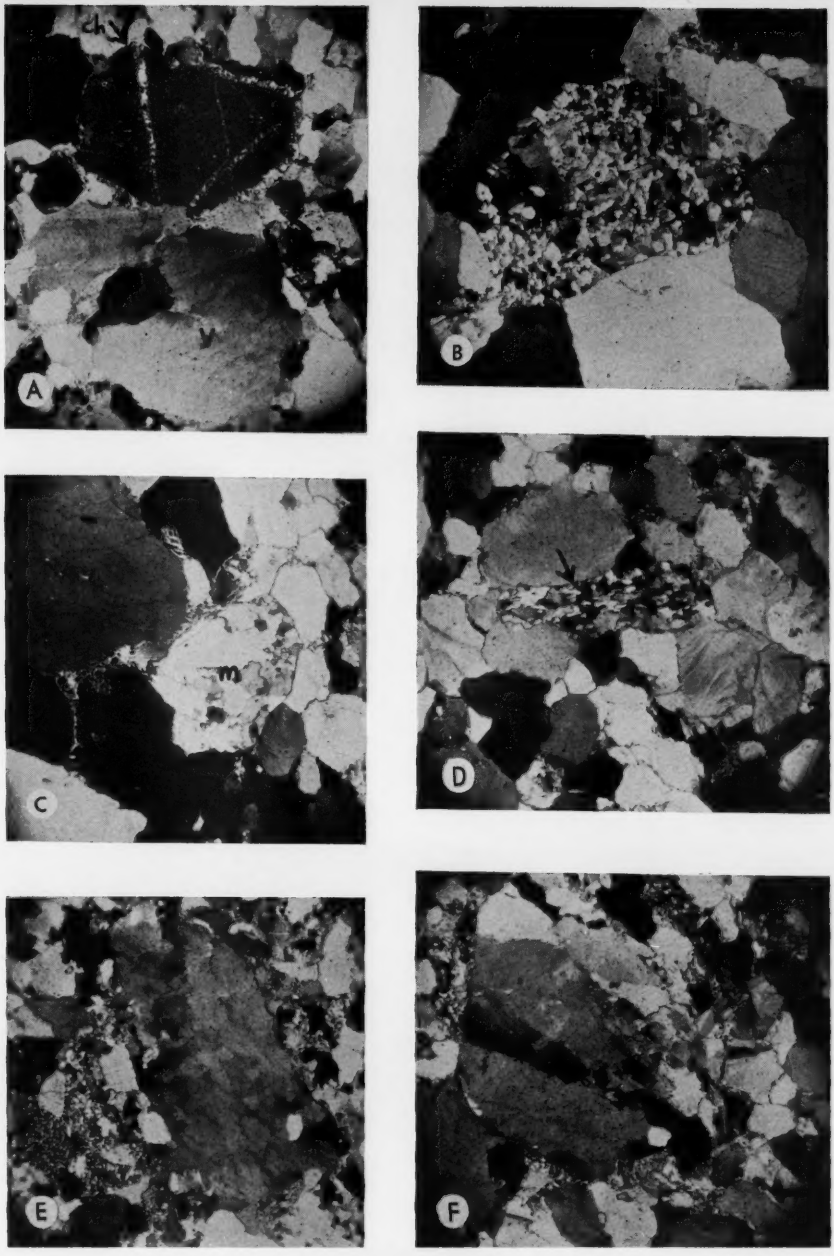

PLATE 3.—Photomicrographs of chert and quartz types, lower part of the upper (white) Tuscarora.

190 ft) contains considerable chert, stretched and composite metaquartzite, but only occasional traces of slate and phyllite fragments; hence it varies compositionally between orthoquartzite and quartzose subgraywacke (plate 2, D, E). All samples from the upper part of the white Tuscarora (from about 190 to 310 ft) are very pure and classed as orthoquartzites (plate 2, B) consisting almost entirely of common quartz with some vein quartz although there is a little metaquartzite near the bottom.

Although these changes within the white Tuscarora are gradational and there is considerable fluctuation, nevertheless the sharpest change (wherein both chert and metaquartzite drop to relatively permanent low levels) occurs at a stratigraphic level of about 190 ft, somewhat less than halfway up the white Tuscarora. This mineralogic break seems to coincide with the top of the basal gravelly zone, and also with a permanent increase in the roundness of the quartz.

The mineralogic zonation is most obviously seen in the hand specimens, where the proportion of chert in the pebble fraction drops off rapidly in going up the section. A census of all the gravel grains larger than 2 mm was made on all the collected samples (total 264 gravel particles) with a hand lens. This showed the following changes: in bed 15 (97.5 ft), 70 percent of the gravel grains are chert, the rest being quartz; beds 16 and 18 (102.5 ft, 120 ft) have 15 to 20 percent chert; by approximately 170 ft, chert comprises only 2 to 15 percent of the pebbles; by 230 ft, chert is down to a trace to 3 percent; and between 250 and 305 ft there is no chert and the pebbles consist entirely of quartz. In the lower zone (below 190 ft) the chert pebbles are notably larger than the quartz while in the upper half of the white Tuscarora the rare chert pebbles appear to be about the same size as the associated quartz pebbles.

The same changes appear in the thin sections. Below 190 ft, chert comprises 2 to 6 percent of the entire rock and is very conspicuous because it forms the largest granules or pebbles in the slides (plate 2, F; plate 3, A). After the sharp decrease at about 190 ft, chert gradually continues to diminish upward. Considering the interval above 190 ft, 2 to 6 *single sand grains* of chert are visible in each slide in the lower part of this section, gradually declining to 0–2 grains per slide in the upper part of the interval. In the upper zone, further, chert grains are the same size as the quartz or even somewhat smaller (.20–.30 mm).

Descriptive mineralogy.—The chief mineral variety in the white Tuscarora is common quartz. Below 190 ft, it comprises about 65 to 85 percent of the rocks; between 190 and 245 ft it forms 80 to 95 percent; and above 245 ft to 305 ft it is almost the sole detrital constituent, forming uniformly 96 to 98 percent of the detritus (plate 2, B; plate 4, B). The relatively low proportion of common quartz in the lower part of the section is caused by the abundance of such materials as chert, metaquartzite, and vein quartz. Furthermore, gravelly samples anywhere in the section have low percentages of common quartz because the maximum size of common quartz is about 1 or 2 mm, and all the granules or pebbles are vein quartz, metaquartzite, or chert. Nevertheless, considering only the sand-size grains, there still is a significant upward increase of common quartz within the sand grains themselves. Originally most of this quartz had straight extinction, but because of late structural deformation (overturned folds) it has been rendered moderately to slightly undulose in most slides. Some quartz shows very pronounced "pinching" effects caused by the impingement of neighboring grains as a result of folding.

Composite quartz grains (made of subequant crystal units lacking undulose extinction and sutured boundaries) show a decrease upward in about the same way as chert. Below 190 ft they form 5 to 12 percent of the slides and are abundant in sand-size grains as well as pebbles, while above 190 ft they comprise a mere 1 to 3 percent. In the Tuscarora there is a complete gradation between composite quartz (derived from recrystallized metaquartzite, plate 3, B, C) and stretched metaquartzite (plate 2, E; plate 3, D); both are made up of .02 to .10 mm quartz mosaic evidently of metamorphic origin. If individual crystal units of the quartz mosaic are elongated, strongly undulose, or lenticular, grains are arbitrarily classed as stretched metaquartzite and

counted as metamorphic rock fragments.

Vein quartz, identified by its characteristic extinction and richness in vacuoles (plate 3, A, E, F; plate 4, A), ordinarily makes up 1 to 5 percent of the slides except in pebbly beds, where it may comprise as much as 10 to 20 percent of the rock (since most of the pebbles are vein quartz). The proportion of vein quartz is fairly constant up the section if one ignores the pebbly beds, and considering both pebbly and non-pebbly beds, the white Tuscarora probably contains 3 to 6 percent vein quartz. All these quartz types appear to have similar roundness in the white Tuscarora.

The decrease in quantity of chert up the section from 2 to 6 percent near the base of the white Tuscarora to virtual absence near the top has already been discussed. In hand specimen the chert pebbles have a dull, chalky white, or cream-colored appearance. In thin section it is seen that some of the chert is pure, homogeneous microcrystalline quartz (plate 3, A), but most of it contains oriented sericite flakes (plate 2, F) or, less commonly, quartz silt grains; these sericitic cherts grade into rocks that are virtually cherty shales, with abundant clays, fine micas, and a little quartz silt in a chert matrix. Most of these cherts and shaly cherts appear to be unmetamorphosed, although some show small quartz veins (plate 3, A) and evidence of minor stretching which probably occurred when the parent sedimentary rock sequence underwent folding. No fossil ghosts were seen in the chert grains. The chert pebbles have rounded corners but tend to a blocky (joint-controlled ?) over-all shape; the chert sand grains are well rounded, especially those in the upper part of the section.

Not one single grain of feldspar was seen in all the slides of the white Tuscarora.

Soft metamorphic rock fragments are rare in the white Tuscarora in contrast with their abundance in the red Tuscarora. They occur only in the lower part (below 220 ft), and even here they are present in less than half of the slides, never forming more than 1 or 2 percent of the specimen. These fragments consist of weakly metamorphosed sericitic clay slate or silty clay slate, most of them smaller than the associated quartz; apparently these are the last remnants that have escaped destruction by abrasion. A few larger grains of very weakly metamorphosed quartzose siltstone (constituent grains .03 mm) also occur. Stretched metaquartzite (plate 2, E; plate 3, D) decreases upward markedly in the same manner as chert and composite quartz; it comprises approximately 3 to 10 percent below 190 ft and decreases to a trace to 3 percent above 190 ft. Some of it has a little sericite, but most of it is nearly pure granulated quartz. Stretching of the quartz units is not very pronounced.

Heavy minerals are scarce. Subround, colorless zircon is most abundant, followed by subround tourmaline, mostly green with some brown. A few tourmaline grains have small overgrowths. Chalky white leucoxene is common, some of it having associated needles of authigenic titanium minerals. A trace of rutile was seen.

Nature of the quartz pebbles.—A special examination was made of the quartz gravel to elicit more details as to the nature of its source. Quartz pebbles and granules are uniformly well rounded, and most of them are milky in hand specimen. Microscopic examination showed that the great majority of the gravel size grains are apparently of hydrothermal vein quartz origin inasmuch as they show the typical semicomposite, blocky extinction (plate 3, A, E) and sometimes comb structure (plate 3, F). They have been weakly squeezed but no badly crushed grains were found. A few of the pebbles consist of composite quartz, either stretched metaquartzite with slightly crenulate boundaries and undulose extinction or recrystallized metaquartzite with individuals having straight extinction. None of them have been badly squashed, however. None of the pebbles are sandstone or even mildly metamorphosed sandstone. Only a few of the smaller gravel-size grains are common quartz, but grains finer than 1 mm are very largely common quartz. Pebbles in the Green Pond Conglomerate of New Jersey (Thomson, 1959) have similar characteristics.

Original vacuole content is difficult to evaluate inasmuch as many of the rocks are crossed by abundant parallel bubble trains imposed by post-depositional structural deformation. But there is a definite tendency for the pebbles and granules to be more

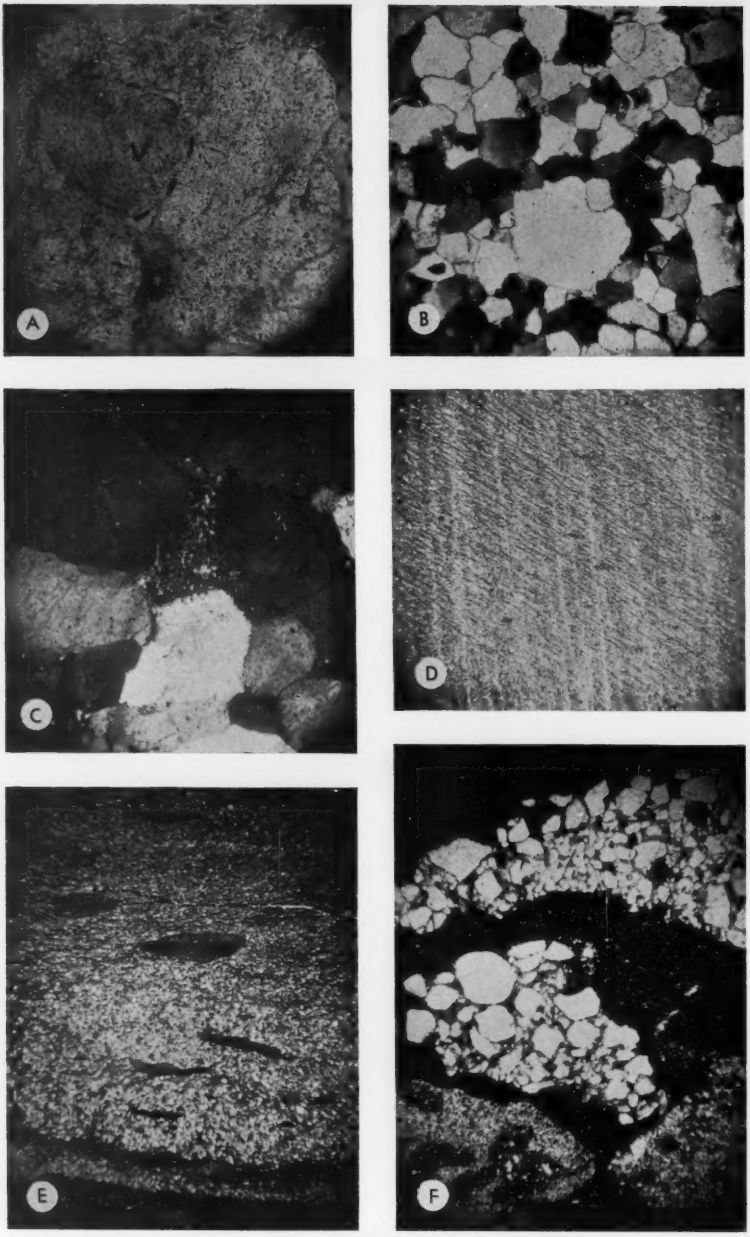

PLATE 4.—Photomicrographs of specimens of Tuscarora and Rose Hill Formations.

bubbly than the sand grains in the same slide, although this is not a sharp difference and there is considerable overlap.

Therefore, it is probable that most of the quartz sand is *not* coming from breakdown and abrasion of the vein quartz pebbles but is from a different source. The prominent bimodality of the Tuscarora (that is, pebbles or granules vs. medium sand) is consequently caused by derivation from two different types of source rock, veins providing the coarser mode, and a more fine-grained source (granite or older sandstones) providing the much more abundant finer mode. Results in close parallel with the Tuscarora situation have been found by Blatt (1958) on New Jersey beaches, where bimodality was caused by a mixture of pebbles of vein quartz with sand grains made of common quartz.

Cementation.—Little or no visible pore space remains in the white Tuscarora. Pore space is lacking for two reasons: (1) in some sands, pressure solution has been an important process, the quartz grains dissolving at points of contact (often producing mildly sutured boundaries) and gradually approaching each other so that pore space originally there was slowly eliminated and the grains now interlock almost in the fashion of a jigsaw puzzle (plate 4, B); (2) in other rocks there has been no pressure solution, and the pore space has been filled with quartz overgrowths forming 5 to 10 percent of the rock. All stages are visible between no pressure solution and abundant quartz cement through partial solution and a small amount of quartz cement to much pressure solution and no visible quartz cement. The quartz occurs as overgrowths, sometimes bounded from the parent grain by a line of vacuoles, but at other times the contact is almost invisible. Normally there are no inclusions in the quartz overgrowths. There seems to be no correlation of stratigraphic position with cementation or degree of pressure solution and no evident correlation with texture of the rocks. Extensive pressure solution has been observed by Thomson (1959) in the Green Pond Conglomerate (Tuscarora equivalent) in New Jersey, and similar relations to the abundance of quartz cement were observed. There pressure solution was promoted by presence of clay films; this does not seem to be a factor in the Tuscarora since clay films are either absent or present only as traces.

There is no direct evidence to the time of origin of the quartz cement although the full-bodied slate fragments in the red Tuscarora indicate probable quartz cementation prior to much compaction. Source of the silica is not known.

Two slides contain .01 to .10 mm elongate idiomorphic crystals of authigenic titanium minerals, probably sphene, anatase, or rutile, protruding from grains of leucoxene and projecting into the quartz cement.

Most specimens below 235 ft contain a trace to 1 percent of thin films of illite between many of the grains of quartz; often the illite is lightly stained with limonite. The illite here appears to be detrital, although the evidence is not clear. It may be

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EXPLANATION OF PLATE 4

A. Grain of vein quartz (v), identified by abundant tiny vacuole inclusions; note cloudiness of grain compared to relative clarity of neighboring grains; Upper (white) Tuscarora. Ordinary light, × 96.

B. Field typical of the supermature orthoquartzites of the upper part of the white Tuscarora. Rock is composed almost entirely of common quartz. Observe interlocking of detrital grains caused by pressure solution. Crossed nicols, ×24.

C. Pore space filled with vermicular books of authigenic kaolin. Upper part of the Tuscarora Formation. Crossed nicols, ×81.

D. Typical olive clay shale of the Rose Hill Formation, consisting of about 95 percent clay and 5 percent visible quartz silt scattered throughout. Thin dark lines are incipient slaty cleavage; lighter streaks runnning vertically are defects in grinding the slide. Ordinary light, ×28.

E. Rose Hill Formation, showing a mud-shale lamina grading upward into nearly pure clay shale. The mud-shale, consisting of nearly equal proportions of silt and clay, has a sharp base; several elongate clay galls are enclosed in the layer. Ordinary light, ×9.6.

F. Specimen from upper part of Rose Hill Shale, showing sand and silt "mottles" embedded in clay matrix. These are probably caused by burrowing organisms. Note striking difference in the grain size between mottles. Ordinary light, ×9.6.

TABLE 1.—*Properties of the three lithologic zones of the Tuscarora Formation, Mills Gap, West Virginia*

Zone	Property						
	Color	Gravel Content	Muddy Matrix	Sorting	Bedding	Rounding[1]	MRF's,[2] Chert, and Metaquartzite
Upper part of white Tuscarora (190′–305′)	White	Small amounts	None	Good to moderate	Semi-random crossbedding	High	All virtually lacking
Lower part of white Tuscarora (95′–190′)	White	Moderately large amounts	None	Moderate to poor	Semi-random crossbedding	Moderate and variable	MRFs almost lacking, metaquartzite and chert common
Red Tuscarora (0′–95′)	Dark red	None	Common	Poor	Crossbedding very common, dip consistent	Nil	MRFs very abundant, metaquartzite and chert common

[1] Amount of rounding being accomplished at the site of deposition (excluding grains with inherited roundness).
[2] MRF—soft metamorphic rock fragments (slate and phyllite).

that the illite came from disaggregation of the slate or phyllite fragments which finally disappear just below this horizon. Between 250 and 285 ft most of the slides contain some pore spaces .05 to .20 mm across filled with books of well-crystallized authigenic kaolin (plate 4, C) and sometimes authigenic, colorless illite. These minerals are much better crystallized and clearer than the detrital clays and also occur as equant masses within former large pore spaces scattered sporadically throughout the rock. It is difficult to tell the relation between these authigenic clay minerals and the authigenic quartz overgrowths. In some parts of the rocks authigenic clays encrust quartz overgrowths; in others, the clays lie on naked quartz grains. In several beds the quartz has grown large overgrowths which are laced with illite shreds arranged in a rectangular latticework pattern. It appears that the quartz and illite had been precipitating at about the same time.

In some samples, post-depositional folding has induced bubble trains that traverse the slide with common orientation, cutting across many detrital grains. In these slides, vein quartz grains can still be recognized by their denser concentration of vacuoles.

General Discussion

All of the Tuscarora Formation in the section studied was laid down in the nearshore zone. However, rates of supply of detritus and wave or current potency at the site of deposition varied considerably so that three successive environmental realms can be distinguished in the formation. Table 1 summarizes the properties of these three zones.

Between the base and top of the formation most rock properties undergo a consistent change toward both textural and mineralogical maturity; clayey matrix disappears, sorting of sand grains improves, rounding increases, and metamorphic rock fragments, metamorphic quartz types, and chert diminish markedly. The biggest change occurs at the boundary of the red and white Tuscarora; it is accompanied by a complete change in the nature of the crossbedding, grain size, textural maturity, and detrital mineralogy. Post-depositional hematite is present in the red and absent in the white Tuscarora, but it must be stressed that the red color developed after deposition by chemical processes and was not produced by erosion of a red regolith on some source land. In addition to the changes caused by shifting environments, the Tuscarora is further complicated by the fact that the source area changed character during deposition of the formation as an extensive sedimentary cover was stripped off to reveal a basement complex.

Depositional environment.—The red Tuscarora shows a very close petrographic affinity with the underlying Ordovician Juniata Formation and serves as a transitional phase between the Juniata and the white Tuscarora. The Juniata is a sequence of red shales and red graywacke sandstones, apparently formed on a series of migrating sluggish river channels and accompanying floodplains, probably near the terminus of a large delta built up at the mouth of an extensive river system (Tuttle, 1940; Swartz, 1948; Krynine, 1946); the Queenston red-

beds of New York represent the distal end of this deltaic complex. The red color indicates an oxidizing environment, probably continental because of the consistent crossbedding and the lack of fossils. Furthermore, the abundance of soft slate fragments indicates an environment where fine and medium sand grains were not being effectively abraded, such as a sluggish river. The Juniata contains very abundant well-rounded quartz grains, but these were inherited as shown by the accompanying sharply angular grains of quartz and soft slate. If quartz rounding were going on during Juniata deposition, the metamorphic rock fragments would have been destroyed and all the quartz grains would have have some edge-rounding.

The Juniata, a fairly clear-cut continental-fluvial formation, is succeeded by the red Tuscarora which is seemingly on the borderline between a continental and near-shore marine environment. The well-developed crossbedding dipping consistently west to northwest, the red color, and the common clay and silt matrix indicate close affinities with Juniata conditions. The mineralogy of the two formations is virtually identical, and the richness in slate fragments and lack of quartz rounding in the present environment (although again many inherited rounded grains are present) indicates an environment like that of the Juniata with gentle currents insufficient to cause effective abrasion. The red Tuscarora *differs* from the Juniata in the absence of interbedded shale section (although red clay galls are common), more massive bedding, and presence of *Arthrophycus* imprints. According to Pelletier (1958), *Arthrophycus* is apparently a "strand line" type of fossil, seeming to flourish right at the transition zone between marine and continental environments, and the data of Amsden (1955) also support this view. Integrating these facts, we can picture the depositional environment as lying athwart a very flat coastline near the mouths of large rivers carrying down sand in a westerly to northwesterly direction across a level coastal plain, occasionally depositing small layers of clay which would get torn up in the next flood to form clay galls or might dry out and be reworked by the next high tide or storm. No permanent deposition of clay beds (as happened in the Juniata floodplains) occurred. Currents were unidirectional but sluggish enough to prevent destruction of the soft slate fragments and not accomplish any rounding of quartz; clay matrix was only partially removed. Thus the area may have been a broad estuarine muddy sand flat, usually inundated in shallow marine water but protected from much surf action. The rate of deposition of the red Tuscarora was probably slower than that of the deltaic Juniata as evidenced by the lack of shale beds, better sorting, and presence of clay galls that indicate reworking. *Arthrophycus* (borings ?) were probably made when this flat was covered at high tide, or perhaps the creature was able to migrate up the river mouths or estuaries a short distance. The water must have been shallow over this flat, and aerated enough to maintain an oxidizing environment for the precipitation of hematite.

As will be shown in a forthcoming part of this research, the redbeds of the Bloomsburg Formation, higher in the section, are interpreted as having formed very near river mouths where the iron was being carried in from the source area and was precipitated upon encountering more saline marine waters. The same explanation is also used for the red sandstones in the Rose Hill Shale (which immediately overlies the Tuscarora). A similar situation may occur in the Tuscarora, with the red muddy Tuscarora being deposited near the estuarine mouths of large rivers carrying in iron, while the white Tuscarora was deposited along beaches more removed from fluvial influence. We should then expect the red Tuscarora to have an erratic lateral distribution and be subject to appearance at various stratigraphic levels in the lower part of the formation (which fits the facts rather well). If the Juniata is considered a subaerial delta, then the delta must have been submerged to produce the littoral Tuscarora sands. Hence we may interpret a minor marine transgression near the end of Juniata delta deposition, which drowned the distributary river systems to produce a series of well-developed estuaries at their mouths. In these, the transitional marine red Tuscarora developed.

The lower part of the white Tuscarora at Mills Gap shows a sudden increase in the potency of winnowing and abrasion by wave or current action combined with an abrupt influx of gravel. A renewed pulse of uplift in the source may have heightened relief and increased stream gradients all along this part of the coastal plain. Then streams, no longer drowned at their mouths, were able to carry gravel the full distance to the shoreline. These fluvially-transported gravels were then worked on by surf action, much as the fluvial sandy gravels of New Jersey are today being attacked along the Atlantic coast (Blatt, 1958). The rapid rate of influx of coarse material and the wide range of sizes contributed did not allow very good sorting to be accomplished. Nevertheless, enough energy was exerted to winnow out all clay and silt, and there is definite evidence that some rounding of medium to fine-grained quartz sand was now taking place at the site of deposition. The red Tuscarora below contains many quartz grains that are quite angular with sharp corners and jagged edges; in the lower part of the white Tuscarora, all the grains have smoothed corners and most of them are subround. Round and well-rounded grains are present in both units, but these have been inherited from older sediments. Every slide of the red Tuscarora shows the same mixed assemblage of rounded and angular quartz, and there is no slide-to-slide variation of roundness, indicating that no rounding was going on at the site of deposition and all the roundness was inherited from sediments in the source area. On the other hand the white Tuscarora shows considerable slide-to-slide variation in roundness, indicating rounding was being accomplished during Tuscarora deposition in some beds more than others (fig. 3).

As further evidence of heightened abrasion, virtually all of the soft slate or low-rank phyllite fragments were eliminated, although abrasion was not sufficient to get rid of the harder chert or composite metaquartzite grains. Clay galls, common in the low-energy estuarine environment, could not be preserved in this regime either. The much more random nature of the cross-bedding combines with all the previous lines of evidence to indicate a probable high-energy littoral environment with shifting currents and waves as the final locus of deposition.

As the source area wore down and gradients decreased, the gravel content declined irregularly. Then in the upper part of the white Tuscarora sorting further improves, rounding attains consistent high degrees, and chert and polycrystalline metamorphic quartz all but disappear. Here all the quartz grains show uniform high roundness, indicating that they were deposited in a stratigraphically persistent environment of potent rounding action and high abrasive effectiveness. Even grains derived from fresh granite or vein sources were able to become well-rounded during one cycle of deposition in the upper white Tuscarora. Together with the nearly random cross-bedding this indicates a high-energy beach environment where influx of detritus had slowed so that the waves and littoral currents had full opportunity to abrade the grains and winnow them. A minor renewed pulse of uplift (?) caused the influx of gravel above 240 ft. Finally, the Rose Hill open marine clay shale transgressed over the last of the Tuscarora beach sands.

Yeakel (1958) believes that the Tuscarora in its outcrop belt is fluvial. In addition to the evidence of potent abrasion given above, the presence of *Arthrophycus* as far east as the Shawangunk Facies in eastern Pennsylvania, argues against this idea.

Source area.—One can postulate most efficiently about source area characteristics during Juniata and red Tuscarora deposition because in these dominantly fluvial or estuarine sands very little abrasion took place in the final depositional environment, and the mineral constituents are retained in close to their original proportions. The high content of weakly metamorphosed slate fragments and various types of metaquartzite indicate that low-rank metamorphics were very important in the source area; however, the rank of the argillaceous rocks was so low that no discrete mica flakes larger than .02 mm had been produced. The abundance of pure chert and clayey or silty chert indicates an older sedimentary source of considerable volume as do the orthoquartzite siltstone fragments and rounded

single grains of quartz silt. At least half of the quartz sand is subround or better and very likely came from an older (Cambrian?) sandstone with rounded grains, probably with a modal diameter of .20 to .25 mm since this size has the best rounded grains. The split round grains also came from older sediments. The source of the angular common quartz is not known; it may have come from older angular sandstones, but it is more likely derived from coarse-grained non-strained metamorphics or plutonic igneous rocks. Only one grain of orthoclase was seen; this could mean a very humid climate and flat topography, in order to cause thorough weathering, but the thick volume of deposits, general deltaic nature and evidence of rapid fluvial deposition favors the interpretation that plutonic igneous rocks were exposed only over a small proportion of the source area during this time.

In summary then, during deposition of the lower Tuscarora the source area was made up of extensive low-rank metamorphics, chiefly slate and metaquartzite, injected by a few quartz veins and with small amounts of granitic rock exposed; these were blanketed by a sequence of rather well-rounded sandstones and chert-bearing clayey limestones. This is in close agreement with the nature of the source area for the Juniata in central Pennsylvania (Tuttle, 1940; Krynine, 1946). Apparently this area was actively folded as shown by the fractures in the chert.

This conclusion agrees with the data of Woodward (1941) that in places Martinsburg Slate is unconformably overlain by Tuscarora, and pebbles of the slate are worked into the basal Tuscarora. The tiny slate fragments so common in the sandstone also probably came in part from the Martinsburg. As Krynine (1946) aptly states, the geosyncline was now autocannibalistic, feeding upon its own weakly metamorphosed flank sediments.

There is a very distinct change in mineralogy between the red and the white Tuscarora. The most important change is the drop in metamorphic rock fragments from 12 to 25 percent in the red Tuscarora to 0 to 2 percent in the lower part of the white Tuscarora to complete absence in the upper white Tuscarora. Nearly all the chert and polycrystalline metamorphic quartz also disappears in passing from the lower to the upper part of the white Tuscarora. There are two possible explanations for this change. One explanation would be that the source area during lower Tuscarora deposition was blanketed by slates, metaquartzites and chert, and as erosion proceeded, these constituents were stripped off until a granitic basement was revealed which yielded nothing but common (plutonic) quartz and a little vein quartz. A second explanation would be that these constituents, which are all less durable than single quartz grains, were eliminated by more severe abrasion in the high-energy beach environment of the upper Tuscarora.

To answer this problem, soft metamorphic rock (slate) fragments do reappear in small quantities (5 percent) in the next two formations up the section, the Rose Hill Shale and Keefer Sandstone. Therefore, assuming that the source area maintained about the same geographic position, if metamorphic rock fragments were present in the source area during Juniata, red Tuscarora, and Rose Hill-Keefer deposition, they should have also been present during deposition of the intervening white Tuscarora. Because they are very soft, they could easily be entirely eliminated by beach abrasion (Todd and Folk, 1957) and yet be present in large quantities in the fluvial or deltaic sands below or the low-energy marine sediments above. Therefore, abrasion is favored as an important cause for the disappearance of slate fragments. But since they never reappear in the Silurian (or Lower Devonian) in amounts anywhere near approaching their abundance in the Juniata or red Tuscarora, a large part of their outcrop must have been eroded through also.

Probably chert to some extent was also selectively removed by abrasion because in sand size grains it wears slightly faster than quartz (Folk and Ward, 1957). This is confirmed by the fact that farther up the Tuscarora section chert grains become not only less abundant but also smaller and better rounded. However, the cover of chert-bearing sediments must have also been stripped off almost entirely by the end of Tuscarora

deposition here because chert never reappears in more than very small traces in any of the higher Silurian or Lower Devonian (Rosenfeld, 1953) formations. This is even true in sands like the Rose Hill, Keefer, or Bloomsburg that consist of angular grains where the chert could not have been removed by abrasion, hence must have been virtually absent in the source. Since chert *pebbles* wear at about the same rate as quartz pebbles (Sneed and Folk, 1958) and the chert pebbles are characteristically subangular and blocky in shape, the change in the ratio of chert to vein quartz pebbles from lower white Tuscarora to upper white Tuscarora is probably more a matter of stripping off the sedimentary veneer from the source rather than selective abrasion.

The reason for the decrease in polycrystalline quartz is unknown. This material also does not reappear in the higher formations in any quantity, even in the angular or coarse-grained sands where it should show up best and would not have been abraded out selectively. It should have nearly the same abrasion resistance as common quartz; therefore, it probably also disappears upward by complete erosion and removal of its outcrop in the source terrane or by permanent drainage changes rather than being selectively eliminated by wear.

Quartz veins must have been abundant in the source area during deposition of the white Tuscarora; presumably the veins underlaid the chert-bearing sediments because vein quartz pebbles increase in proportion up the section as chert decreases.

Inasmuch as chert, polycrystalline quartz, and metamorphic rock fragments virtually disappear in the upper part of the white Tuscarora, this part of the section is made up almost entirely of well-rounded common quartz with small amounts of hydrothermal quartz. The common quartz could have come from three possible sources: (1) nonstrained, recrystallized metamorphic sources (2) older sandstones, or (3) granite. A recrystallized metamorphic source is not likely because, as shown before, the fine-grained recrystallized metaquartzite disappears rapidly upward (and virtually none is present in the sands of the overlying Rose Hill or Keefer); it is very difficult to conceive of a metamorphic source that would contribute only single crystal units with no composites and no associated finely crystalline metaquartzite.

Deciding between a granitic source and an older sandstone (or sandy limestone) for the common quartz is virtually impossible from direct evidence in these superbly rounded beach sands. In the Juniata and red Tuscarora, however, this question can be answered quite easily. In these fluvial or estuarine formations no quartz rounding took place during deposition, and the grains are preserved in essentially the same shapes they had when liberated from their parent rocks. Thus there is a mixture of angular grains of probable igneous or unstrained metamorphic derivation together with subequal quantities of well-rounded sand grains coming from older sedimentary deposits; the two shape populations are quite clearly differentiated. In the lower part of the white Tuscarora some rounding did take place in the depositional environment, enough to blunt the edges of the angular grains. But many of the high-roundness inherited grains can still be recognized, although it is now impossible to set any quantitative value on their proportion. It appears to be very roughly similar to the quantity in the lower (red) Tuscarora; this deduction is aided by the fact that both the red and lower part of the white Tuscarora have similar chert content, the chert and rounded quartz presumably coming from the same general sedimentary terrane. However, in the upper part of the white Tuscarora the intense beach abrasion was able to round all the grains to a high degree, and no longer is there any shape distinction between the grains that started out as angular fragments derived directly from a granite and those that started as rounded sedimentary grains. In examination of the medium-grained quartz sands higher in the section (rare grains in the Rose Hill red sandstones, and common ones in the Keefer Sandstone) it is evident that by far the greatest quantity of these grains is angular or subangular, and there is very little contribution anymore from well-rounded older sandstones. This is even true in the lowest red sandstone of the Rose Hill, 120 ft above the top of the Tuscarora in which all the medium and coarse sand grains are subangular. Thus the

rounded sandstone source was contributing abundantly during Juniata and red Tuscarora deposition and had virtually ceased supplying detritus in the overlying Rose Hill and Keefer. The upper part of the white Tuscarora should then show a gradual decline upward of the sandstone source to negligible amounts in the top of the Tuscarora. Consequently, the bulk of the upper white Tuscarora was derived directly from granite and quartz grains became well-rounded in one cycle of deposition (fig. 8).

Integrating this information, we may visualize the source area during deposition of the red Tuscarora as consisting of two dominant terranes: (1) a large area consisting chiefly of weakly metamorphosed clay slates with associated hydrothermal quartz veins and belts of metaquartzite presumably overlain by (2) a series of folded sedimentary rocks, chiefly chert-bearing clayey limestones plus supermature orthoquartzite sandstones of fine to medium grain size and orthoquartzite siltstones. A few small areas of granitic rocks may have projected through the metamorphics. As erosion proceeded to produce the large volume of detritus during Tuscarora deposition, the source was uplifted, and gradients increased so as to carry in gravel. Gradually much of the sedimentary cover was stripped off, in particular the cherty limestones, and the metamorphic area was also breached to reveal increasing quantities of granitic rock. During upper Tuscarora deposition, the source probably had been denuded so that it consisted mostly of granitic basement although belts of slate and sandstone still existed in many places. Uplift ceased and the area was worn down to lower relief so that all feldspar was eliminated by a combination of humid climate, slow mass movement of detritus, and heightened abrasion in the beach environment of the upper Tuscarora. Subsidence of the basin of deposition caused widespread transgression, and further quiescence in the source brought an end to the Tuscarora sands and initiated deposition of the Rose Hill marine clay on top of the beaches.

It is interesting to note that the modal grain size of the quartz is almost identical in the fluvial Juniata, fluvial-estuarine red Tuscarora, and littoral white Tuscarora. Because the quartz grain size is independent of environment, this indicates that the source area is the main control on particle size; this must have been the modal size of the quartz grains in the parent rocks whether that rock was an older granite (during upper Tuscarora deposition) or an older sandstone that was in turn derived from granitic rocks (as in the lower Tuscarora). Although the source area controls the modal grain size, the final environment of deposition is responsible for eliminating or retaining the mud fraction of the detritus, abrading out the softer constituents, and rounding the quartz (but without significantly changing its size). The detritus was carried from the source rocks to the depositional site by rivers which accomplished virtually no abrasion of sand-size quartz as shown by the fact that all non-inherited grains are angular in the lower Tuscarora and Juniata.

Climate.—There is no positive evidence of climate during Tuscarora deposition. The probable exposure of extensive granitic terrane during upper Tuscarora deposition together with the total lack of detrital feldspar would seem to favor a warm humid climate for its complete removal. The meaning of authigenic kaolin in the pore spaces is not known, but this could indicate a lot of dissolved silica and aluminum compounds derived from weathering in the source; a large influx of silica from this source would have produced the abundant quartz overgrowths. The hematite in the red Tuscarora implies deep weathering of ferruginous minerals in the source. Better evidence shows that the overlying formation, the Rose Hill Shale, was deposited during a period of rather warm and humid climate also.

ROSE HILL SHALE

Regional Relations

The Rose Hill Shale was named by Swartz and others (1923) after exposures of olive and minor purplish shale at Rose Hill near Cumberland, Maryland. It is regarded as equivalent to the lower part of the Clinton Group of western New York and represents a marine transgression following the coarse detrital wave of the Tuscarora. The Taconic uplift had now spent itself, and presumably

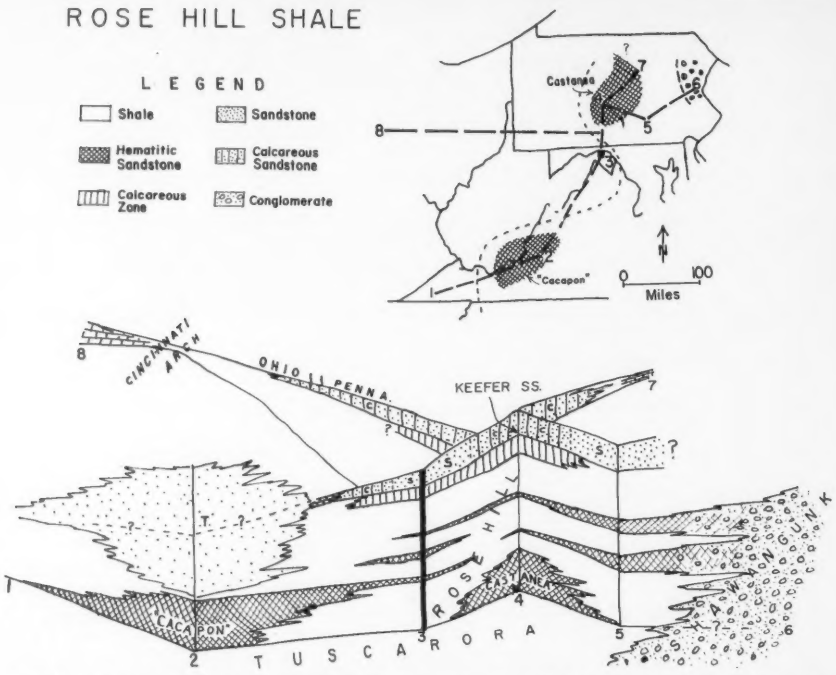

FIG. 4.—Regional variation in the Rose Hill Shale (diagrammatic). Section studied for this report is labeled "3" on both fence diagram and map and is 540 ft thick. Observe local masses of hematitic sandstone ("Cacapon" and Castanea); the section studied apparently lay in a deeper marine clayey embayment during Rose Hill deposition. The dashed line on this map represents a hypothetical depth contour; the shoreline lay to the east some distance (perhaps 100 miles) but was probably of similar form with a deep embayment opposite panhandle West Virginia. In the fence diagram the connection of the thin red sandstones in section 3 with the "Cacapon," Castanea, or Shawangunk Facies is purely diagrammatic; none of them have actually been traced into these masses. Note passage of Rose Hill into Shawangunk Conglomerate in eastern Pennsylvania and into thick Tuscarora-like sandstone (T) in section 2, south-central Virginia.

This fence diagram also shows relationships of the Keefer Sandstone; for a map of Keefer Facies see figure 7. The Keefer thins away from central Pennsylvania where it is fossiliferous and calcite-cemented (c), and becomes quartz-cemented and non-fossiliferous to the southeast (s). Like the Rose Hill it is abruptly replaced by the thick sandstone mass of section 2.

the source area was of subdued relief. Throughout most of its outcrops the Rose Hill consists very largely of olive-colored argillaceous shale, but in some areas purplish shale beds are common; thin siltstone or very fine sandstone layers are characteristically scattered throughout in small quantities. Beds of oolitic or fossiliferous hematite (Alling, 1947), which were formerly worked for ore in New York, central Pennsylvania, and extreme southwestern Virginia are present locally; these Clinton hematite beds are still the mainstay of iron ore production for the Birmingham, Alabama, steel industry. Furthermore the Rose Hill is characterized by the presence of large but geographically limited thicknesses of deep red hematite-cemented quartz sandstone (5–20 percent hematite) which occur in different places at different stratigraphic levels. Figure 4 shows the regional relationships of the formation.

The Rose Hill is thickest along an axis from south-central Pennsylvania through

western Maryland and central West Virginia, thickness along the axis gradually decreasing southward from 800 ft in central Pennsylvania (including the basal Castanea red sandstone) to 600 ft in western Maryland and 400 ft in southwestern West Virginia (Woodward, 1941).

In central Pennsylvania the Rose Hill consists largely of olive shale with some purplish shale and very thin hematite beds, but there is up to 75 ft of very hard, deep red hematitic sandstone at the base, designated by Swartz (1934) as the Castanea. This sandstone contains *scolithus* tubes, worm castings, and rare ostracods and thins rapidly to the south and east. In many places the Castanea is transitional to the underlying Tuscarora (Swartz, 1939) and could be considered to be an upper member of that formation. The rest of the Rose Hill here contains a few 1 to 3 in layers of hematitic sandstone, but these thicken considerably into eastern Pennsylvania, and much of the Rose Hill passes farther eastward into the upper part of the Shawangunk conglomeratic mass (Swartz and Swartz, 1931). A similar sequence is found in New York where shale, limestone, and hematite in western New York pass into sandstone and then into part of the Oneida Conglomerate toward the east (Alling, 1947). In central Pennsylvania, the upper 50 ft of the Rose Hill contains thin beds of fossiliferous limestone (Swartz, 1934). This calcareous zone continues across Maryland into northern West Virginia but disappears into eastern Pennsylvania. In the type area near Cumberland, Maryland, the formation is chiefly olive shale with some purple shale near the top, and a 10 to 30 ft thick mass of red iron sandstone (Cresaptown) 175 ft above the base (Swartz and others, 1923).

In northwestern Virginia and the panhandle of West Virginia the Rose Hill consists very largely of olive shale with little or no purple shale, and there is no thick red sandstone at the base although a few very thin red sandstones occur at several levels. Woodward (1941) divides the formation into three parts: (1) the upper part which is most fossiliferous and contains the only limestone beds in the Rose Hill, sporadic beds only a few inches thick. (It is this upper calcareous zone that extends north into central Pennsylvania, but dies out southward into central West Virginia.); (2) a middle portion, mostly olive-green shale with very sparse fossils; and (3) the lower Rose Hill, mainly greenish, moderately fossiliferous shale. In west-central Virginia a marked change occurs. In the lower half of the Rose Hill more and more iron sandstones become intercalated with the shale until they dominate the section between Botetourt County, Virginia, and Monroe County, southern West Virginia. These red sandstones, designated the "Cacapon" Facies (Butts, 1940), contain a few ostracods and clay galls. In this area the upper Rose Hill also appears to be replaced by a thick mass of white quartz sandstone, possibly an expanded Keefer. Passing farther into southern Virginia, the red "Cacapon" Sandstone decreases again as the Rose Hill resumes its ordinary shaly character, but it begins to pick up more hematite beds and purple shales (Miller and Fuller, 1954); traced still farther southwest into Alabama, the Clinton consists of sandstone (some of it rather coarse), varicolored shale, and the hematite ore beds that supply Birmingham. Passing west from the Appalachians, the Rose Hill thins and consists very largely of shale; west of the Cincinnati Arch it grades into limestone.

Thus the Rose Hill shows a complex facies pattern. In northern West Virginia (the area of study) the lower part of the formation is largely olive shale, but this is replaced by thick localized masses of red iron sandstone in central Pennsylvania (Castanea) and in west central Virginia ("Cacapon"). Furthermore, West Virginia has almost no beds of ore-grade hematite, but these are more common in central Pennsylvania and west central and southwest Virginia and reach great importance in northeast Alabama. Purple shales also seem to be associated with these high-iron areas. As shown later, the red sandstones are believed to represent marine but shallow, nearshore environments; the same is probably true of the oolitic and fossiliferous hematite ores (Alling, 1947). Thus it can be interpreted that the northern West Virginia area studied in this report represented a farther offshore (probably also deeper) marine embayment bordered by shallower near-shore facies in central Pennsylvania

and west central Virginia; the shore line must have made a large eastward bulge across the present outcrop belt (fig. 4). This may reflect the presence of local uplifts in the source area in New York, New Jersey, and south-central Virginia which resulted in the build-up of extensive sheets of coarser red detritus while no such uplift occurred in that part of the source lying directly east or southeast of the West Virginia panhandle. The local southern uplift may have been active also during Tuscarora deposition since the Tuscarora is conglomeratic in Mercer County, West Virginia and during Keefer deposition since the Keefer thickens enormously in the same locality as the "Cacapon" sandstones appear.

These postulated uplifts were apparently sporadic in place and time, because in middle and upper Rose Hill the red sandstones apparently indicate a source in eastern Pennsylvania. During uppermost Rose Hill deposition clay influx was seemingly slowing in the central Pennsylvania-northern West Virginia area since occasional thin limestone beds first appear.

The fauna of the Rose Hill consists chiefly of ostracods which enable an excellent zonation (Swartz, 1934, 1942), but brachiopods are also common and there are some trilobites and other fossils. These occur in the shale and also rarely in the iron sandstones.

The section studied for this project is about 30 miles southeast of the axis of maximum thickness and is interpreted as lying near the northern side of the marine embayment; it is about 80 miles south of the well-developed Castanea and 150 to 200 miles north of the "Cacapon" iron sandstone. Thus the section is nearly all olive clay shale with a few thin intercalated gray siltstones. There are only a few thin red sandstones (mostly in the middle of the section) probably representing attenuated tongues of the larger masses to the north and southwest; there are no beds of iron ore.

Local Character

An excellent section of the Rose Hill Formation is exposed along the Baltimore and Ohio Railroad tracks ½ to 1 mile southwest of the old station at Sir Johns Run, 2 miles northwest of Berkeley Springs, Bath Township, northern Morgan County, West Virginia. Here the beds strike approximately N 36° E and dip 55° NW; minor drag folding and some minor faulting makes the thickness estimates somewhat inaccurate. This writer measured a thickness here of 540 ft, although Woodward (1941, p. 62) measured 438 ft at this locality. Samples were taken at regular 15 ft intervals, with extra samples collected where warranted.

The Rose Hill Shale consists of 90 to 95 percent of dull olive-colored, rather hard clay shale of high purity. Within this great thickness of shale occur interbedded siltstones and sandstones of three distinct types which enable the formation to be divided roughly into three zones although the predominant shale is similar throughout the formation (fig. 5).

The lowest zone, 0 to 300 ft, consists predominantly of olive clay shale, but intercalated with this are thin beds of a very hard, blue-gray well-sorted siltstone cemented with ankerite and quartz. The siltstone forms lenticular beds ½ to 1 or occasionally 3 in thick spaced from one to several ft apart within the monotonous shale section. In the Sir Johns Run section these siltstones are virtually absent in the lowest 80 ft of the formation, but between 80 and 140 ft, gray siltstone beds form 15 to 25 percent of the section. From this point to 300 ft the siltstone beds form only 2 to 5 percent of the formation, but they increase again just before the first red sandstone at the beginning of the middle zone of the Rose Hill Shale. One 2-ft bed of hematitic siltstone is present at the 110 ft level, within that part of the lowest zone that is especially rich in gray siltstone beds.

The middle zone, 300 to 390 ft, likewise consists of about 90 percent olive clay shale, but the distinguishing rock type of this zone is a hard purplish-red sandstone cemented with hematite and quartz (fig. 5). These sandstones form 5 to 10 percent of the section and occur in beds ranging from 6 in to 2 or 3 ft thick. Occasional thin beds of hard gray siltstone like that in the zone below are intercalated with the olive shale and red sandstone. This zone appears to be somewhat above that of the Cresaptown iron sandstone of Cumberland, Maryland some

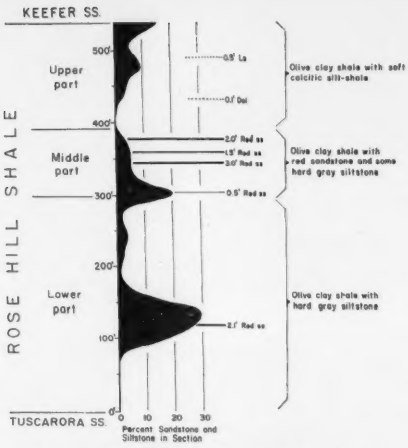

FIG. 5.—Zonation of Rose Hill Shale section at Sir Johns Run, Morgan County, West Virginia. In all zones, olive clay shale is greatly predominant, forming approximately 90 per cent of the rocks; the shaded area shows proportion of sandstone or siltstone units in the section. It is based on a moving average covering 20 ft of section; for example, the 20-ft section centered at 250 ft (that is, the interval 240 to 260 ft) contains 3 percent sandstone or siltstone and 97 percent olive clay shale. Positions of red hematitic sandstones are shown by heavy horizontal lines; they are prevalent in the middle zone. Siltstone beds in the lower zone are very hard and cemented with quartz and ankerite; those in the upper zone are soft, fissile, micaceous, clayey, and often cemented with calcite. Two thin carbonate beds appear in the upper zone which is the most calcareous and fossiliferous part of the Rose Hill in northern West Virginia and central Pennsylvania (see fig. 4).

30 miles to the west (Swartz and others, 1923).

The upper zone, 390 to 540 ft, averages 95 percent olive shale. Here the shale is interbedded with a lighter gray siltstone of a different type than that encountered below. These siltstones occur in beds 1 to 3 in thick but are much softer and less well cemented than the siltstones below in the section; some are somewhat friable and porous. Instead of having a massive, smoothly conchoidal fracture they split into thin plates a few mm thick and should be termed siltshales because of these partings. They contain more mica and clay than the previous siltstones and, unlike the others, effervesce in acid because of calcite cement. Nearly all the siltstones above 390 ft are of this type, but a few "normal" hard gray siliceous siltstones reappear near the top of the section. Within this upper zone of the Rose Hill the quantity of siltstone increases gradually from about 1 percent near the base of the zone to 10 percent near the top of the section. Two very thin limestone beds were also found in this zone.

In detail, the four dominant rock types of the Rose Hill have the characteristics discussed below.

Olive Clay Shale

This rock type forms 90 percent of the Rose Hill Shale and is of quite constant character throughout the formation. The clay shale units are normally 1 to 5 ft or so thick separated by the thin interbeds of sharply-differentiated siltstone or sandstone layers as described above; however, some sections of shale continue uninterruptedly for 20 ft or more. The shale breaks into slabs 2 to 5 mm thick and is moderately hard. Normally no texture or color banding or lamination is visible on crossbreaks (except for fissility), and ordinarily no silt or sand is detectable with the hand lens so that the material is exceedingly homogeneous clay. The color is remarkably constant pale grayish olive, averaging 10Y5/2 although some specimens range between 5Y to 5GY, 4/2 to 6/2. Specimens have a faint sheen on the bedding surfaces because of the excellent orientation of the constituent illite or sericite. Some beds are crossed by weak slaty cleavage making angles of up to 40° with the bedding. The color is due to a mixture of the greenish illite, sericite, and chlorite with tiny limonite specks. Weathered surfaces become coated with a thin rich brownish film (5YR4/4, moderate brown).

Microscopic characteristics.—The great majority of specimens of this rock type are classed as clay shale. Some thin ($\frac{1}{8}$- to $\frac{1}{4}$-in) laminae of mud shale (subequal amounts of silt and clay) are encountered in thin section or by careful study of the hand specimens, but these form no more than 15 to 20 percent of this rock type. Rare specimens contain nests, balls, and irregular contorted stringers of silt and sand (see below) mixed with the dominant clay.

The typical clay shales (plate 4, D) average about 5 percent of visible quartz silt and 95 percent clay; the silt content ranges between 0.1 and 10 percent. Quartz silt occurs as grains averaging .015 mm and ranging from .005 to .030 mm; these grains are randomly scattered throughout the clay as isolated single grains. This indicates a lack of current action; probably the silt grains were windblown and simply settled at random into the clay mass. In the less common mud-shales silt may form 30 to 50 percent of the rock and average .020 to .040 mm. Presence of mud shale laminae implies weak current action; some of these grade upward into clay shale by decrease in the size and quantity of the silt (plate 4, E). The clays have a very perfect physical and optical orientation (Z parallel with bedding), and, when present, mica flakes are lined up perfectly with the bedding (not the slaty cleavage), indicating that the sediment was not disturbed by currents, wave action, or burrowing organisms after deposition. In some specimens with well-developed slaty cleavage the clays may be oriented at some angle to the bedding.

Terrigenous minerals form over 98 percent of this rock type. Clays, of course, are predominant and appear to be chiefly illite, chlorite, and sericite under the microscope inasmuch as they are pale greenish; their index is higher than balsam and they give a good creamy to yellowish birefringence. The length of the flakes appears to average about 5 μ, but they range up to 10 or rarely even 20 μ in length. The latter may represent packets of closely-oriented flakes. Mica (chiefly muscovite) forms an exceedingly small proportion of the clay shales, averaging about 0.1 to 0.2 percent. The mica flakes average about .05 mm long, and there is a very distinct gap at about .02 to .03 mm between the discrete mica flakes and the mass of clay surrounding them. The quartz silt grains, averaging .015 mm, are subangular and subequant and generally show undulose extinction even in these very small grains. Since the quartz grains are suspended like raisins in raisin bread throughout the yielding clay mass, this strain must have been induced in the source area; that is, the quartz has not been strained by post-depositional folding of the shale.

It is interesting to note that no fossils or carbonate grains were noticed in thin section; furthermore, the rocks appear to be quite low in organic matter. The only noticeable orthochemical constituent is 1 to 5 percent of yellow-brown, earthy limonite occurring in distinct scattered grains 5 to 20 μ in diameter. The shapes of these tiny grains are difficult to make out, but many appear to be cubic and probably represent pyrite pseudomorphs.

Six typical Rose Hill olive clay shale specimens were studied by X-ray diffraction to determine the clay mineralogy more precisely. These samples were run by Mario Messina, and the patterns were interpreted by Edward C. Jonas. Samples were spaced equally up the stratigraphic section and were very similar except for quartz content. A powder pattern was run on the entire shale sample, and an oriented-flake pattern was run on the minus 2 μ fraction for each sample. Samples were heated to 600° and re-run to check presence of kaolinite. The predominant clay mineral is well-crystallized illite or sericite; there are also major amounts of chlorite and kaolinite. The chlorite peak nearly disappears in the $<2\ \mu$ fraction; thus the chlorite flakes are probably larger than illite or kaolinite. Quartz content was roughly estimated by Jonas as of the order of 10 percent, which corresponds favorably with the average petrographic estimate of 5 percent visible quartz.

Clay shales with silt or sand nests.—Occasionally clay shale specimens contain nests, patches, lumps, balls, or contorted stringers of silt or sand obviously much coarser than the rock surrounding them. The sand or silt in these patches is well sorted in general (although there is much clay interstitial to the grains) and may range in median from .05 to .15 mm. In one remarkable specimen near the top of the section there are nests of sand, each nest fairly well sorted within itself although the median in different nests may vary from 0.1 to 0.5 mm, and there are even some granule-size particles floating in the clay matrix (plate 4, F). Some of the sand balls have a concentric "snowball" structure but most have none; some have been strung out or distorted on folding in incipient boudinage. The origin of these peculiar balls is not known; they may represent coprolites in some instances but more

probably are "mottles" caused by burrowing animals (Moore and Scruton, 1957). Some of the mud-shale layers contain 1 to 4 mm clay galls (plate 4, E; plate 5, A), indicating that the currents that formed these coarser streaks were strong enough to erode the clay beds below.

Hard Gray Siltstone

This rock type occurs chiefly in the lower zone of the Rose Hill but occurs sporadically throughout the formation. The beds are ½ to 2 or 3 in thick, somewhat lenticular, and very sharply differentiated from the olive clay shale in which they lie. Specimens show very faint even horizontal lamination on a scale of 1 to 2 mm; occasionally faint gently dipping cross lamination is observed. The beds are extremely hard and non-porous, usually breaking with a smooth conchoidal fracture and presenting an almost flint-like appearance because of the quartz cement. The color when fresh is medium gray with a faint blue-green tinge, 5BG5/1, caused apparently by the illite and chlorite content and the ankerite cement. Weathered specimens show a several-mm-thick dark brown rim (5YR2/2) often accompanied by diffusion banding. The weathered portion is slightly porous and friable and only a little less hard than the fresh rock. Apparently the only change that takes place is the alteration of ankerite to limonite in place; the brown color is not caused by a pervading stain but simply by the change of each ankerite crystal to a limonite pseudomorph. Fresh ankerite in the interior of the rock gives way sharply to ankerite which has thin films of limonite entering along cleavage planes and crystal margins; gradually the limonite spreads out into the crystal and near the edge of the rock the whole crystal is made up of earthy brown oxide. No change apparently takes place in the chlorite or the quartz cement on weathering. It is interesting to note the depth of limonite alteration in a rock apparently as impermeable as these siliceous siltstones.

Microscopic characterististics.—A typical specimen is described as a well-sorted very fine sandy siltstone: siliceous, ankeritic, mature orthoquartzite or subgraywacke. Considering grain size, most of these rocks are classed as well- to moderately-sorted unimodal very fine sandy siltstones (plate 5, B) with a few silty very fine sandstones and some pure siltstones. They average approximately 3 percent clay (ranging from 0 to 10 percent) and the modal grain size varies from .04 to .07 mm. Sorting in terms of σ ranges from .4ϕ to .6ϕ, and a typical sample has a median of .055 mm with two-thirds of the grains lying between .035 and .07 mm. In nearly all specimens the size of the very largest grain in the thin section ranges between .10 and .15 mm, but one slide of silty very fine sandstone had a secondary mode at .20 mm (comprising only a fraction of one percent) and also contained one .80 mm quartz grain. Physical orientation of occasional mica flakes and elongated quartz grains parallel with the lamination is prominent, and in all slides the slow-ray vibration direction of the quartz grain lies in the plane of the bedding and parallel with the long dimension of the grains. This is readily tested by insertion of the gypsum plate; the quartz grains look dominantly blue if the bedding is oriented in a NE-SW direction and dominantly yellow at right angles to this. Further evidence of current action is the presence of occasional laminae rich in heavy minerals and layers of slightly coarser or finer silt.

Terrigenous minerals comprise about 75 percent of this rock type. Compositionally, the rock hovers on the borderline between orthoquartzite and subgraywacke, averaging about 1 percent feldspar, 2 to 7 percent metamorphic constituents, and 92 to 97 percent quartz. Most quartz grains are subangular to angular, although a few subround grains .04 to .06 mm occur. There are a fair number of sliver-like quartz grains but most are subequant. From 1/10 to ½ of the quartz grains show weak undulose extinction, but to some extent this is the result of post-depositional folding; some grains contain rather abundant vacuoles and are apparently from hydrothermal sources. Chert grains form about 0.2 percent of the detritus in the lowest siltstone specimen examined, but in the higher ones only minute traces were seen. Moderately fresh subround orthoclase without overgrowths averages approximately 1 percent; both chert and feldspar are the same grain size as the quartz. Mica averages 0.1 to 0.5 percent (the most micaceous specimens containing as much as 1 to 2 percent) and forms thin flakes

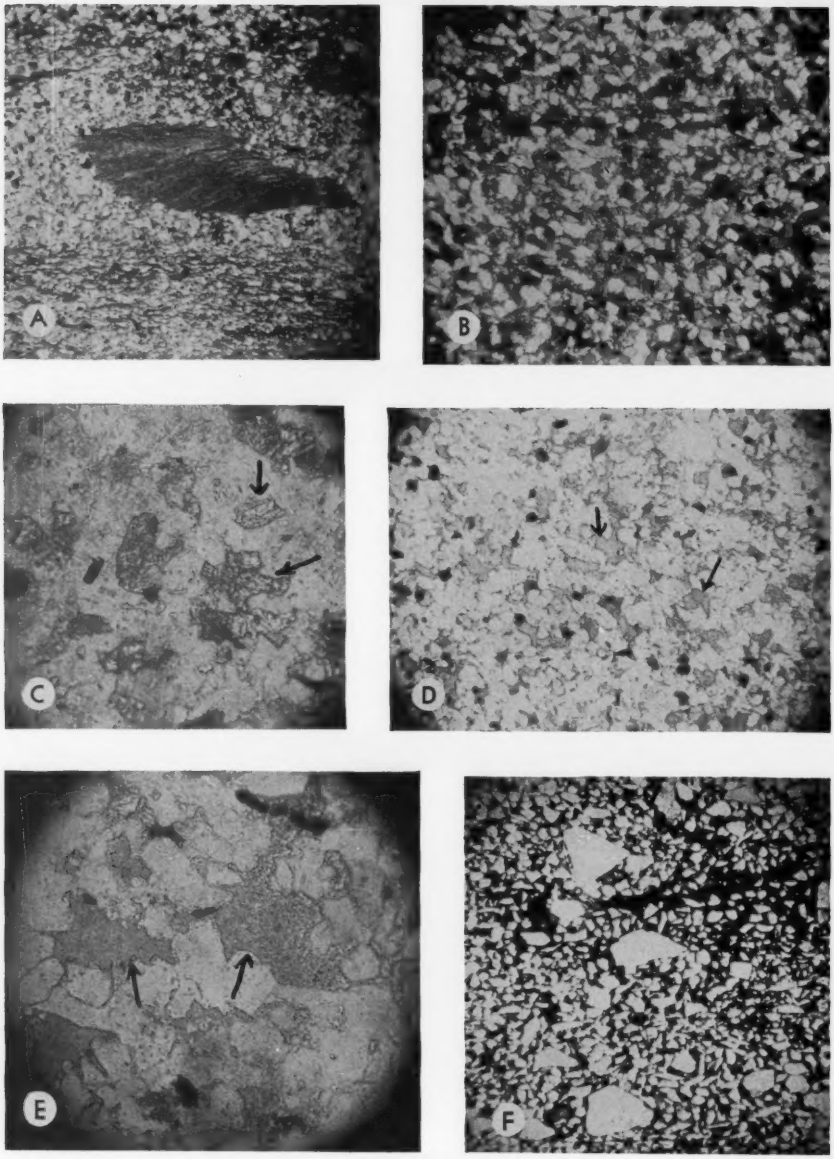

PLATE 5.—Photomicrographs of Rose Hill Formation.

averaging .10 mm and ranging up to .20 or .30 mm in some slides. Definitely terrigenous mica in the form of distinct, thin plates consists of approximately 85 percent muscovite, 10 percent chlorite, and 5 percent fresh rich brown biotite. Metamorphic rock fragments comprise 2 to 7 percent of the grains; usually they consist of finely crystalline stretched composite quartz (microcrystals averaging about 15 to 20 μ), but in some slides slate fragments are common. Slate fragments are particularly abundant in a siltstone sample at 300 ft, just below the zone of red sandstones which are also richer in slate fragments. The slates consist of finely recrystallized quartz and sericite with some chlorite. Illitic clay forms thin discontinuous coatings around the sand grains in some slides. Much the most abundant heavy mineral is leucoxene which in some slides forms 1 to 2 percent of the rock as dull, earthy chalk-white rounded to subangular grains. In occasional slides some of the micas are also permeated with leucoxene stain. Subround to subangular grains of rich orange-brown, clear green and blue gray tourmaline and colorless, subangular zircon are common as heavy minerals, but magnetite is absent except for very small amounts occurring at the base of the section.

The most prominent orthochemical mineral of this rock type is ankerite which averages 15 to 25 percent of the specimens (plate 5, C). It is identified as ankerite because of its alteration to limonite and by its refractive indices. It occurs as uniformly distributed single grains averaging .05 mm in diameter usually in the shape of subequant xenomorphic to hypidiomorphic rhombs. Grains are of fairly constant size and shape and appear to have formed before quartz cement for two reasons: (1) they touch detrital silt grains preventing formation of quartz overgrowths, and (2) they appear to be the same size as the surrounding silt grains, indicating that they have probably grown by pushing the surrounding unconsolidated silt grains aside (pore spaces between silt grains on deposition should not be as large as the grains themselves). Quartz occurs as almost invisible overgrowths on silt grains which makes them appear interlocked in a mosaic. Quartz is estimated to form 5 to 10 percent of the rock and fills all available remaining pore space. Occasional small grains of pyrite occur, and in two slides authigenic chlorite patches .05 to .30 mm in diameter form 15 to 30 percent of the rock; these are the only two specimens that lack ankerite cement. Perhaps the iron went into chlorite instead of ankerite. The chlorite patches appear superficially like pore-fillings and are composed entirely of typical green, Berlin-blue birefringent, microcrystalline chlorite with the microcrystals randomly oriented and averaging 3 to 7 μ (plate 5, D and E). Sometimes it occurs between the quartz silt grain and its quartz overgrowth but more commonly occurs on top of the overgrowth. The chlorite formed before tectonic deformation because it is cut by joints. It may be a replacement of glauconite or may be a direct precipitate. The antipathetic relation with ankerite cement seems to indicate a competition of these two minerals for the available iron. If the chlorite is a replacement of a mechanically-deposited iron mineral like glauconite, this

EXPLANATION OF PLATE 5

A. Clay gall in mud-shale lamina; ordinary light, ×25.

B. Photograph showing typical good sorting and fine grain size of the texturally mature hard gray siltstone beds in the Rose Hill. Observe subequant shapes of the silt grains and good grain orientation. Average grain size of this specimen is .055 mm. Crossed nicols, ×25.

C. Ankerite crystals (high relief) in a well-sorted hard gray siltstone. The rock is also tightly cemented with quartz overgrowths. Ordinary light, × 100.

D. Chlorite patches (medium gray) in the hard gray siltstone of the Rose Hill. Black grains are heavy minerals. Observe good sorting in this coarse siltstone. Ordinary light, ×29.

E. Same as D, detail of large chlorite patches. Ordinary light, ×100.

F. Hard red hematitic sandstone, middle zone of the Rose Hill. Most of the grains are well-sorted very fine sand, but there is a prominent secondary mode consisting of much coarser subangular quartz. Many highly elongate quartz slivers occur among the finer grains. Grain orientation is good. Ordinary light, ×10.

might indicate that the glauconite (?) used up any iron that ordinarily went to form ankerite in the non-glauconitic (?) normal siltstones. One of the siltstones with chlorite pellets is cemented with 30 percent calcite instead of ankerite.

Hard Red Sandstone

This rock type occurs chiefly in the middle zone of the Rose Hill Formation, between 300 and 390 ft, where it comprises 5 to 10 percent of the interval in beds ranging from 6 in to 2 or 3 ft thick. In this interval it is intercalated between thick beds of olive clay shale and occasional thin hard gray siltstone beds. A single bed of hard red sandstone also occurs at a level 110 ft from the base of the Rose Hill. Red sandstones occur in those parts of the section where the ratio (siltstone plus sandstone):(clay shale) is the highest; that is, measuring up section one finds a higher than normal proportion of gray siltstones as opposed to clay shale, and this heralds the appearance of a red sandstone farther up. Beds show in thin section faint, horizontal laminae about 1 to 5 mm apart. The beds are extremely hard and nonporous, some breaking with a conchoidal almost glassy fracture because of the quartz cement; others with much hematite and little quartz have a hackly, earthy fracture. The color is a dull, deep purplish red, 5R to 5RP3/1, because of the presence of hematite. The faint purplish cast may be caused by the green of chlorite blending with the red of the hematite. Specimens weather with a thin brownish film covering the surface.

Microscopic characteristics.—A typical specimen is described as a moderately sorted silty fine sandstone: siliceous, hematitic submature chlorite-bearing subgraywacke to orthoquartzite. These rocks vary texturally from silty very fine sandstone to fine sandstone. The modal grain size ranges from .075 to .15 mm, and they contain 0 to 2 percent clay. Sorting is good to moderate, ranging from $\sigma = 0.4\phi$ to 0.7ϕ; two of the samples had secondary modes in the 0.40 to 1.0 mm range comprising 1 to 5 percent of each slide. These stray coarse grains are quite distinct and float embedded in a sea of very fine sand (plate 5, F; plate 6, A). Grains are quite angular as a rule although many grains in the .06 to .10 mm size range are subround to round, evidently inherited from older sediments. Even the grains of coarse sand in the secondary mode are just as angular, if not more so, than the smaller grains. Quartz sphericity averages rather low with a high proportion of sliver-like quartz grains which are elongated as much as 3:1 or even 8:1 (plate 6, A). These slivers average about twice as long as the modal grains in a given sandstone and probably represent disc-shaped grains in three dimensions. As in the gray siltstones, the elongated grains are well-oriented parallel with the bedding with their slow-ray vibration direction also dominantly paralleling the bedding. Occasional placered heavy minerals are present, and several specimens contain irregular red clay pockets, clay galls, or thin hematitic clay laminae.

Terrigenous minerals form about 80 percent of this rock type. The rocks are classed as borderline between subgraywacke and orthoquartzite, averaging zero percent feldspar, 1 to 15 percent metamorphic rock fragments, and 85 to 95 percent quartz. About one-third of the quartz grains show undulose extinction; this is partially the result of late structural deformation, and partly original. Some of the coarser grains appear to come from veins because they have a moderately large amount of bubbles but most are straight-extinguishing common quartz. Only a few small grains of microcrystalline chert were noted in all the slides, and no feldspar was seen. Flaky micas are extremely scarce because of the relatively coarse size of the rocks, but chlorite pellets form 0.5 to 4 percent of these rocks in subequant round to subangular grains .05 to .10 mm in diameter. The chlorite grains show no evidence of squashing, indentation, or compaction. They are composed of clear green, transparent chloritic aggregate, with a microcrystalline (5–20 μ) or sometimes fanlike aggregate birefringence in low first-order gray or faint bluish gray to nearly isotropic. Occasionally the constituent flakes are oriented, but most commonly they lie at random with the pellet. The origin of these peculiar grains is not certain. Their subangular shape and occasional placering suggests that they are replacements of some heavy mineral such as hornblende; but they are too abundant for this and they may

have formed as primary chlorite pellets or as a chloritic replacement of glauconite. It is not believed that they are fragments of terrigenous chlorite schist because of their subangular shape and the fact that they consist of pure chlorite with no quartz, and the chlorite flecks are not oriented as they are in the metamorphic rock fragments.

Metamorphic rock fragments vary widely from 1 to 15 percent, averaging 5 percent. In the red sandstones they show a complete range from chlorite-sericite phyllite or slate particles consisting almost entirely of mica through all mixtures of finely recrystallized quartz, chlorite, and sericite to pure finely recrystallized metaquartzite. In the gray siltstones, metaquartzite is by far the dominant rock fragment observed; thus the red sandstones contain relatively more of the micaceous fragments although the latter are still subordinate to the quartzose ones. The metamorphic rock fragments are the same size as the quartz in the same slide, but are usually subelongate and distinctly better rounded because of their softness. The fine red sandstones contain more of the softer micaceous rock fragments while the coarser red sandstones have mainly the more durable metaquartzite. Each fragment is made up of microcrystals of quartz averaging 5 to 20 μ (sometimes with sutured boundaries and strong undulose extinction) mixed with parallel-oriented flakes of chiefly chlorite, muscovite, or sericite 10 to 20 μ long.

Leucoxene and magnetite are the most common heavy minerals. Some zircon and tourmaline are present. Leucoxene forms 0.2 to 1.5 percent of the rocks; it forms rounded chalky white grains about .05 mm in diameter which are placered with the other heavy minerals, showing detrital origin. None of the leucoxene grains are tinged with limonite and none show black metallic cores. Magnetite is present in abundance in the red sandstones but is virtually though not completely absent in the other sandstones and siltstones. Grains are round to subround, subequant, average .03 to .06 mm, and are distinctly placered. They are black with a metallic luster and show no sign of alteration; thus they are not ilmenite.

Two specimens (at 120 and 305 ft) contained a few thin arcuate shell fragments 1 to 1.5 mm long totally replaced by quartz and chlorite. They appear to be brachiopod or ostracod fragments.

Hematite and quartz are the chief cements with a very small amount of calcite occasionally present. Where present, calcite was the first cement to develop since quartz grains within the small calcite patches have no hematite rims and no quartz overgrowths, yet the outside of the calcite patch is coated with hematite. Hematite was precipitated from solution following the calcite (if any), and it now forms 5 to 30 percent of the rocks (plate 5, F; plate 6, A). Where most abundant, hematite fills all the original pore space and prevented formation of any quartz overgrowths; where less abundant, it coats sand grains under quartz overgrowths and also occurs as solid masses on top of quartz overgrowths, indicating two stages of precipitation of iron. In many places there is not any hematite at points of contact between quartz grains, further demonstrating that the hematite is post-depositional. The hematite appears earthy in hand specimens and under high power is seen to be made of a myriad of microcrystalline flecks. It is pure hematite free of included clay but shows no tendency to develop oolitic, colloform, or fibrous crusts. Quartz was usually the last cement to form, and filled all the remaining pore space. It comprises 1 to 10 percent of the rocks; where quartz overgrowths are abundant, the rock appears glassy with a conchoidal fracture; where they are almost absent, the hand specimen is dull and earthy hematitic with a hackly fracture. Although in general the paragenetic sequence was calcite-hematite-quartz, there was considerable overlapping. Because the soft chlorite pellets are full-bodied, this indicates that the sands were rigidified by quartz and hematite cementation prior to much compaction.

Specimens are crossed by very thin quartz-filled tectonic veinlets which cut all chemical cements; some veinlets also contain a little chlorite.

Greenish Silt-Shale

This rock type occurs interbedded with olive clay shale in the upper part of the section. Beds are a few inches thick, and

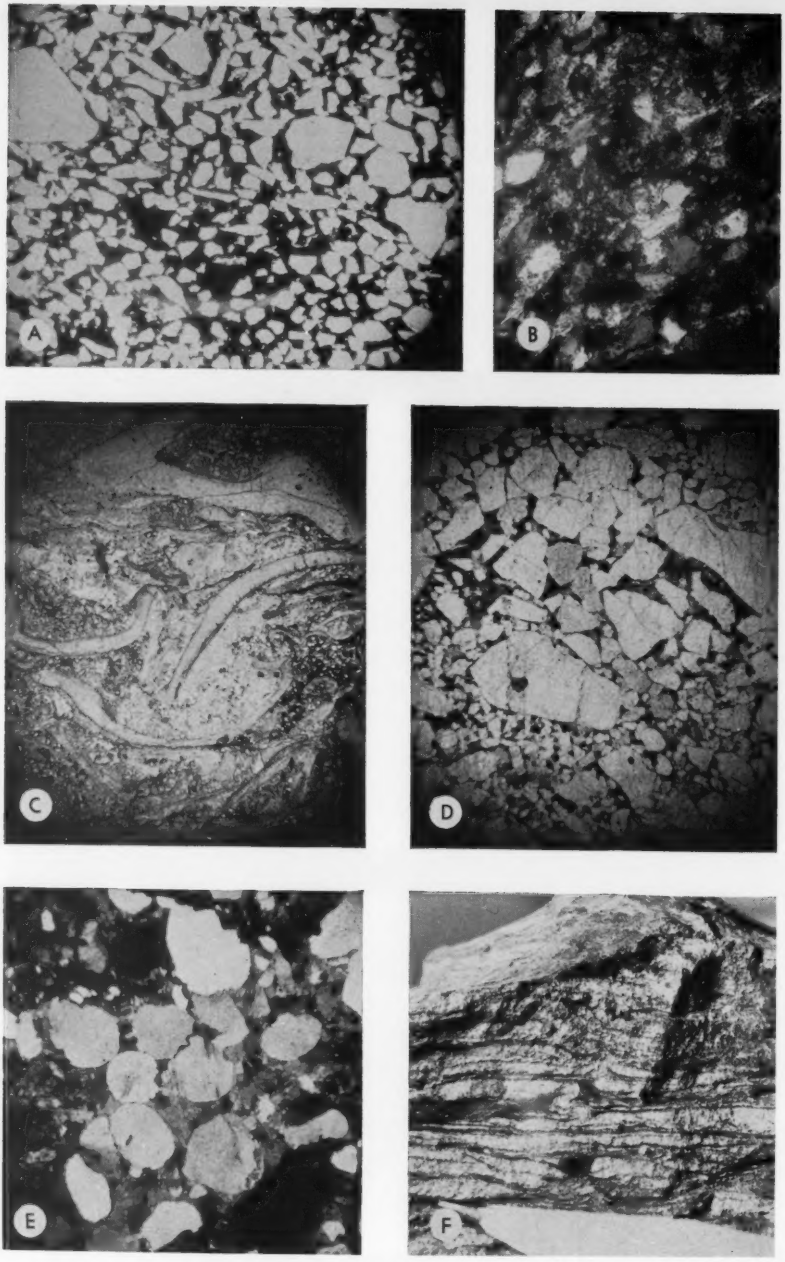

PLATE 6.—Photographs, Rose Hill and Keefer Formations.

are finely laminated with partings every 1 to 5 mm along which they split easily into platy slabs. Mica is richly concentrated along the partings. Usually the laminae (and partings) are gently curved and cross-stratified at a small angle to the bedding plane. These beds are of only average hardness (in contrast with the gray and red siltstones which are extremely hard) and may even be somewhat friable; because of their ease of splitting and relative softness they are termed silt-shales instead of siltstones, using the term "shale" in a purely structural sense. These field differences are due to the much higher content of clay and well-oriented mica and to the near lack of quartz cement. The color is light gray with a faint greenish tinge (5G6/1 to N5, greenish gray to medium gray) paler than the gray siltstones below it; bedding planes become covered with a rich blackish-purple, almost iridescent coating on weathering.

Microscopic characteristics.—A typical specimen is classed as a silt-shale or a mud-shale: immature micaceous subgraywacke (plate 6, B). Most of the specimens are classified as silt-shales or mud-shales; some contain a small quantity of very fine sand. They contain between 15 and 50 percent clay and the remainder is silt with a modal size averaging .030 to .040 mm. The size of the coarsest observed grains varies from .10 to .15 mm, and in one sample grains of this size form a very sparse secondary mode. Silt grains are subangular to subround with only a few sliver-like grains. Quartz grains and mica flakes are very conspicuously oriented parallel with the bedding; many thin laminae are almost entirely mica. Effective porosity prior to cementation was very slight because of the abundant clay matrix.

Terrigenous minerals form 60 to 95 percent of this rock type; recomputing these to 100 percent gives less than 1 percent orthoclase, 1 to 3 percent discrete mica, 2 to 8 percent metamorphic rock fragments (chiefly soft slate and phyllite), and 90 to 95 percent quartz. Thus the rocks are micaceous subgraywackes, barely over the border from orthoquartzites. About one-fifth of the quartz grains show undulose extinction. Mica is much more abundant in these rocks compared with the rest of the Rose Hill, forming flakes averaging .05 to .10 mm and ranging up to .25 mm; muscovite forms about three-fifths of the flakes with chlorite making up the bulk of the remainder and brown biotite about one-twentieth of the mica. Fine-grained fragments of slate or phyllite consisting of recrystallized quartz mosaic with considerable sericite, illite, or chlorite (all pale green) occur abundantly in rounded grains. Leucoxene is the most abundant heavy mineral, with zircon and tourmaline next in

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EXPLANATION OF PLATE 6

A. Hard red sandstone, Rose Hill Formation, cemented with hematite (black). Grains are quite angular, and very elongate quartz slivers are abundant. Ordinary light, ×29.

B. Greenish silt-shale from the upper Rose Hill. This specimen has an illitic-chloritic clay matrix and is rich in mica flakes and slate fragments; thus it is classed as an immature subgraywacke. Grains are well-oriented. Crossed nicols, ×84.

C. Limestone bed (trilobite biosparrudite, that is, coarse trilobite fragments in sparry calcite cement) in the upper Rose Hill. Owing to the irregular shape of the fossil fragments, the packing is very loose and there is much sparry calcite. Ordinary light, ×9.9.

D. Specimen of the lower part of the Keefer Sandstone, showing textural heterogeneity. Quartz grains of every size from 4 mm granules through sand to silt are embedded in a gray clay matrix, thus the rock is a texturally immature orthoquartzite. Some of the large grains are rounded, but most are subangular. Ordinary light, ×9.9.

E. Supermature orthoquartzite bed in the lower zone of the Keefer, well rounded and well-sorted grains of coarse quartz sand in dolomite cement. Almost all grains are common quartz in contrast to the Tuscarora wherein vein quartz and metaquartzite abound. Compare rounding of these grains with those of the same size in D. This indicates that the rounding occurred during Keefer deposition. Grains are somewhat sutured due to solution under pressure. Crossed nicols, ×8.4

F. Specimen from the lower zone of the Keefer, showing lenticular silt stringers and "mottles" in black shale matrix. Observe small scale "recumbent folding" in silt layers. Hand specimen photograph, ×1.5.

abundance. Pale greenish clay (illite, sericite and chlorite) occurs thickly throughout the rock as a matrix.

In one specimen three dolomitized fossils occurred, two brachiopod specimens and one crinoid.

About half of the specimens contain finely crystalline calcite cement in scattered to clustered grains averaging .04 mm and comprising 30 to 40 percent of the rock. Grains are xenomorphic irregular and appear to have grown by pushing the silt and clay aside, as "swimming" terrigenous grains appear in the middle of some large carbonate patches. The rest of the samples contain no calcite.

In the calcite-cemented rocks there is little or no quartz cement. In the non-calcitic ones there are some inconspicuous quartz overgrowths, but generally quartz cement is scarce in these rocks probably because of the abundant clay matrix. Traces of pyrite occur.

Carbonate Rocks

Only two carbonate beds were found, both in the upper part of the formation that is sparsely calcareous as far as central Pennsylvania.

A 1-in bed at 430 ft is a light gray medium crystalline fossiliferous dolomite, containing stringers of very fine sand and silt. Dolomite averages 0.2 mm, but shows a wide size variation with some crystals as large as .7 mm; crystals are unusually xenomorphic for dolomite and have curved cleavages. They have formed by replacement of a biogenic limestone as a few ostracod (?) and crinoid ghosts remain. There is a trace of collophane.

The other bed at 490 ft is 6 in thick and is a trilobite biosparrudite (Folk, 1959a) consisting of about 40 percent fossils (chiefly trilobites with some ostracods, crinoids, and one bryozoan), mostly with large calcite overgrowths (plate 6, C). Fossils are mostly whole with little wear and are 0.5 to 3 mm long. Finely crystalline to sub-fibrous calcite, averaging .05 to .1 mm, forms the cement and has definitely pushed some of the fossils aside during crystal growth. On etching, some lacy chert was found to occur around the fossils. There was no dolomite in this specimen.

General Discussion

In considering all the sediments of the Rose Hill, one of the most striking features is the clean-cut textural differentiation between silt and clay. Nearly all of the units of the formation are either practically pure clay shales (with less than 5 percent of .01 to .02 mm quartz silt) or pure well-sorted siltstones with very little clay; contacts between these two types of lithology are hairline-sharp. Some mud-shales and fine silt-shales in the upper third of the formation do show subequal mixtures of silt and clay, but these rocks are volumetrically minor. If one plots for each specimen the percentage of silt in the rock against the modal grain size of the silt, a compact linear trend develops (fig. 6). This shows that once the modal size of the silt exceeds .045 mm there is little or no clay; but as the silt mode becomes finer, the rocks contain less and less silt (and more and more clay), following the equation: (Clay Percentage) = 50(ϕ mode of silt) − 210. This close relation, and the predominance of either pure siltstone or pure clay shale, seems to indicate that the Rose Hill depositional environment was one of considerable efficiency in transportative differentiation, one whose currents though gentle were able to cleanly separate these two quite fine size fractions into texturally mature sediments. The formation is considered to be less mature in the upper part where clay and silt is intermingled to form the mud-shales.

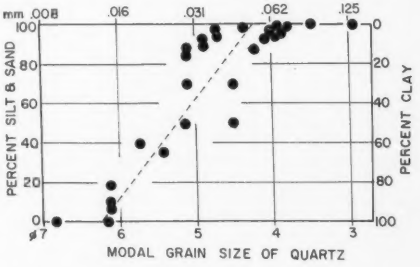

FIG. 6.—Relation between modal grain size of quartz silt or sand and proportion of clay in the Rose Hill samples. These two properties are closely related in the undisturbed marine beds of that formation, indicating effectiveness of currents in accomplishing sorting.

If one were to compile a total grain size curve for the entire Rose Hill, it is apparent that the frequency curve would have two peaks, by far the predominant one in the clay sizes with the second peak in the size range of .03 to .10 mm. There is an apparent lack of material between .01 and .02 mm; presumably this is due to a lack of such fine silts (6 to 7ϕ) in the source area. A similar lack of material in the vicinity of 7ϕ has been reported by Nienaber (1958) and Elliott (1958) in near-shore marine and bay sediments of the Texas coast.

This gap in the size distribution may be characteristic for marine (perhaps for all) sediments, because of the fundamental difference in the origin of quartz sand and coarse silt particles as opposed to clay particles. The former originate by liberation of the quartz grains of such rocks as granites, schists, and phyllites; the quartz within these parent rocks normally ranges from medium or coarse sand to coarse silt size. Some breakage occurs during transport, but it is probably minor so that quartz of a grain size below 10 or 20 μ is relatively scarce. The clays on the other hand originate from chemical weathering or else from physical break-up of slates or phyllites which were formerly shales that had also derived their clays ultimately from chemical weathering. Most clays are finer than approximately 2 μ although some flakes of sericite and fine micas may be up to 10 or more microns long. Thus the size range between 10 and 20 μ is too fine for most quartz produced by nature and also is too coarse for most clay. This is probably the reason for the gap in the size distribution of the Rose Hill Shale.

The differentiation into clean silt against "clean" clay layers is a function not only of current action but also depends on the size distribution gap. Folk and Ward (1957, p. 11) in a study of a gravel-sand river bar found that there was a pronounced size gap in the vicinity of 0ϕ; thus sediments either consisted largely of gravel with not much sand or sand with virtually no gravel; there were few intermediate mixtures of these two size fractions. Stronger currents could carry all the gravel particles, and weaker currents would carry only sand; there was an intermediate range of current strengths for which there were no available particles, hence the sharp differentiation into gravel-rich and pure-sand beds. The same principle holds for most of the Rose Hill although this consists of mixtures of coarse silt with clay. There is a range of current strengths here for which very few particles are available (10 to 20 μ); stronger currents will carry almost pure silt, and weaker ones can carry only pure clay. If the source had provided a complete size range of materials, there would have been no such clean-cut differentiation into clay beds and siltstone beds, rather all intermediate grades would have been present. Such is apparently the case in the upper Rose Hill where mudstones are of frequent occurrence.

Comparing the three types of silty or sandy units, the hard red sandstones in the middle of the section are coarsest, and many of them even contain a minor mode in the medium to coarse sand grade; the mud-shales or silt-shales in the upper part of the formation are finest and contain much admixed clay. In addition to grain size, the three coarser lithologies also differ in detrital mineralogy, grain shape, and type of chemical cement. To a considerable extent the differences in detrital mineralogy and grain shape are a direct function of grain size while the chemical differences though not directly a function of grain size are probably linked to the same background cause—a change in depositional environment.

The most obvious detrital difference is the highly micaceous nature of the mud-shales and the silt-shales compared with the very impressive paucity of micas in the hard gray siltstones and red sandstones and their near absence in the clay shales. This is partly a matter of hydraulic ratio, because the .05 to .10 mm mica flakes apparently are deposited primarily in those rocks consisting of .03 to .04 mm quartz silt embedded in an abundant clay matrix. Units either coarser or finer than this contain very little mica. But partly it is due to a stratigraphic change too, because rocks of the same grain size contain more mica in the upper Rose Hill than those in the lower Rose Hill.

Soft metamorphic rock fragments (slates and phyllites) are scarce in the hard gray siltstones (lower Rose Hill), somewhat more

abundant in the red sandstones (middle Rose Hill), and most abundant in the greenish-mud-shales or silt-shales (upper Rose Hill). This is evidently a stratigraphic change, not associated with grain size differences, and would be in harmony with the upward increase in discrete mica flakes.

The second size-linked parameter is grain shape. Roundness is completely the inverse of normal relationships, since the finest rocks (the .03 to .04 mm silt- or mud-shales) contain the largest proportion of round or subround quartz grains, and the coarsest rocks (the red sandstones) contain the most sharply angular quartz. Even the scattered grains of medium and coarse sand are angular. These facts indicate that no rounding was going on during deposition in the Rose Hill environment, and that the finer round silts were inherited from older sediments. This reworked sedimentary source is confirmed by the presence of occasional chert grains.

The Rose Hill contains a high proportion of quartz slivers, that is, quartz grains more elongate than 3:1 with some as long as 8:1. The slivers commonly range from .10 to .30 mm long and .05 to .10 mm wide, and they tend to have the apparent slow ray vibration direction subparallel to their length. A special study was made of the stratigraphic variation in sliver content, and it at first appeared that they were most common in the middle Rose Hill with a moderate number in the lower part of the formation and few in the upper part. But more detailed consideration showed that the proportion of slivers was almost if not entirely a function of grain size since they were especially prevalent in rocks with a modal size between .06 and .12 mm and much less common in rocks either coarser or finer than this. Hence they seemed to be most abundant in the middle Rose Hill because most sandstones in this interval (that is, the red sandstones) had modes in the high-sliver range; finer grained specimens within this same stratigraphic interval had few slivers.

Chemical differences between the several rock types chiefly involve the role of iron. In the hard gray siltstones the iron is fairly abundant as ankerite cement and as sporadic chlorite pellets. In the hard red sandstones iron is very abundant as detrital magnetite, hematite cement, and chlorite pellets. In the greenish mud- and silt-shales there is very little iron, and the main chemical cement is calcite. There is also little iron in the olive clay shales, and what iron there is takes the form of pyrite.

The abundance and chemical state of iron in the different rock types of the Rose Hill seems to depend on the inferred distance from shore or depth of water. The two units which are finest-grained and therefore probably farthest from shore, the olive clay shale and the greenish silt-shale, both have very little iron in them and it is in a reduced form, pyrite, which agrees well with more stagnant conditions. The gray siltstones, which are coarser and much better sorted than the silt-shales, are probably deposited at intermediate distances from shore; thus they have a moderate quantity of iron, which, under mildly reducing conditions, chiefly went to form ankerite (which contains ferrous iron). The red siltstones and sandstones, which are the coarsest units in the formation, contain most iron, and it is in the oxidized form of hematite; this would fit well with the idea that these have formed in the shallowest, most oxygenated waters nearest the shoreline. The two units that under this idea seem to be nearest the shoreline also contain iron in the form of chlorite pellets. These may have been deposited originally as primary chlorite pellets, or they may be a replacement of glauconite or one of the other sedimentary-iron minerals, but at least they testify further as to the abundance of iron in the depositional environment. The Wills Creek Formation, farther up section, shows a similar apparent richness of iron in nearshore environments, and the iron is in a higher oxidation state the nearer one gets to the inferred shoreline.

The large percentage of magnetite in the red sandstones and its virtual absence in all the gray siltstones (even when they are intercalated with the red ones) is also a chemical phenomenon and does not imply different source areas. Magnetite is stable in an oxidizing environment and therefore persists as fresh grains in the hematitic sandstones. In the gray siltstones deposited under a reducing environment the detrital magnetite was dissolved out. Miller and Folk (1955) have shown that in many redbeds magnetite grains are abundant in the red portions of the rock but are dissolved out

in the gray or green reduced portions, even if that reduced portion be a tiny reduction spot no more than a millimeter or so across.

Depositional environment.—The great bulk of the Rose Hill Formation was laid down in one environment—that which deposited the olive clay shale which forms 90 percent of the total thickness. The monotonous rule of this environment was punctuated from time to time by incursions of coarser beds, representing sporadic increases in current strength and changes in chemical environment. Judging from the over-all abundance of clay, these sediments must have been laid down offshore from a coastal plain threaded with large, turbid rivers.

The environment that predominated throughout the deposition of the formation and formed the olive clay shales must have been one of exceedingly calm currents, because these strata normally contain more than 95 percent clay and less than 5 percent of fine silt. The fact that the silt is scattered in random grains instead of in laminae further indicates lack of currents; the silt may have been blown into the area by wind. The excellent orientation of the clay minerals shows that there was very little disturbance by boring animals (except in a few beds where silt "mottles" occur), and the lack of fossils supports the idea of a sea bottom largely devoid of visible organisms. Although ostracods occur throughout the Rose Hill, they must be discovered by patiently splitting shale beds; volumetrically they are so scarce that none were seen in any of the Rose Hill Shale thin sections. Chemically the environment was midly reducing as shown by the content of a small amount of pyrite. The green color of the shale illustrates, however, that any organic matter present was removed (probably by scavenging bacteria) so that the environment was not reducing enough to be toxic, and black shales did not develop. In summary, the bulk of the Rose Hill was laid down in a relatively calm, moderately deep-water, open-marine, mildly reducing environment with little macroscopic life.

This proposed environment is surprisingly similar to that found by Van Andel and Postma (1954) for the modern Gulf of Paria. This Gulf, which lies between Trinidad and the Orinoco delta, is brackish to normal marine, depending on the season, and is about 50 to 100 ft deep. It is largely floored by sand-free clay that is almost devoid of living megafauna; below a thin surficial oxidized layer, its color is olive green (5GY4/1 to 5Y4/1). These properties are strikingly coincidental with the virtually fossil-free, pale grayish olive, almost pure clay of the Rose Hill, similarly interpreted as a marine mudflat off a coastal plain threaded with large rivers. Van Andel and Postma record the Eh in the main part of the bay as neutral and the pH as 8.0 so that little pyrite or organic matter is present and the colors are not dark. The green color they ascribe to the natural color of illite and montmorillonite, following Keller (1953). These oceanographic properties could well represent the situation during most of Rose Hill deposition.

The hard gray siltstone with its good sorting, general lack of clay matrix, good grain orientation, delicate lamination, and placered heavy minerals illustrates that this calm, offshore environment was occasionally disturbed by gentle, persistent currents. These disturbances happened after 2 to 4 ft of clay had been laid down (probably 5 to 10 ft prior to compaction), and were sufficient to form lenticular beds 1 to 3 in thick. Occasionally these currents possessed enough erosive power to tear up small fragments of the olive clay and incorporate them as clay galls in the siltstone beds. The usual lack of graded bedding and presence of sharp contacts at the top of siltstone beds, presence of only very small amounts of entrapped clay, universal delicate 1-mm lamination of alternate finer and coarser layers, occasional cross-bedding, and lenticularity of the beds, favor deposition by some means other than turbidity currents. The rather abrupt pinchout and lenticular thickness changes of many of the siltstone beds give especially strong evidence against turbidity current deposition since beds laid down by this mechanism should be very persistent laterally and show no pinch-and-swell. Presumably the silt beds could have been formed by normal (low-viscosity) marine currents occasionally impinging into the calm clay-depositing area because of storms, floods, or seasonal changes, or these beds may have formed at times of marine regression when the shoreline was moved temporarily nearer, the water became shal-

lower, and silt could be carried into the environment. The beds must have been laid down below the zone of vigorous wave action and in an area free of burrowers in order for the delicate lamination to be preserved. The currents that brought in the silt also carried considerable dissolved iron which was precipitated as chlorite (or glauconite ?) pellets and later as ankerite cement. The silt in the siltstone beds was a new contribution from the source; it is much coarser than the scattered fine silt grains that lie isolated in the clay shales. Therefore it could not be an accumulation of lag silt grains that had been "combed out" of the clay shales.

The hard red sandstone was seemingly deposited under still more shallow-water marine conditions in a zone of fairly strong currents. This is shown by the lack of clay matrix, good sorting, coarser grain size, delicate lamination, good grain orientation, and placered heavy minerals. Current action was nevertheless not sufficient to accomplish any quartz rounding. The red sandstones show no evidence of continental origin, inasmuch as beds a few inches thick are intercalated between the marine clay shales; furthermore, two of the red sandstone beds contained fossils, probably brachiopod fragments, and fossils, worm castings, and hematite oolites occur in the red sandstones of Virginia and Pennsylvania. Presumably these beds were deposited in a nearshore marine environment near rivers carrying large amounts of iron in solution in addition to their load of silt and fine sand. In this shallow, agitated marine environment, chlorite or glauconite pellets formed, and later hematite cement developed owing to easy access of oxygenated waters. The fact that the shallow-water red sandstones occur in the middle of the formation indicates that there was probably a minor uplift or regression during deposition of the middle Rose Hill in panhandle West Virginia so that tongues of red sandstone were able to spread into this area from the larger masses in central Pennsylvania, Virginia, or from an independent source to the east. The red sandstones are presaged by a significant increase in the quantity of gray siltstones, which favors the idea of an approaching shoreline.

The soft greenish silt-shales and mud-shales of the upper Rose Hill present a problem in environmental interpretation. These sediments are the finest and most poorly sorted of the three silty lithologies and are the only ones that contain abundant clay matrix. Moreover, these sediments are also the only ones that contain an abundance of fine mica and metamorphic rock fragments; preceding silty or sandy beds lacked material of a size between 10 and 20 μ, hence the siltstones were cleanly differentiated from the clay shales. This rock type, on the contrary, is flooded with micaceous material of this intermediate size; thus the textural differentiation and sorting is much poorer. Consequently it is not known how much of the unusual textural character of these beds is caused by a more micaceous source terrane and how much is caused by environmental differences. These rocks may have been deposited in an environment of more gentle currents favored by their very fine grain size and clay content. Yet this environment was capable of producing fine lamination, mica-placering, good small-scale cross-bedding, and excellent grain orientation. Perhaps the best solution is to consider these as lying farther from shore (not necessarily in deeper water) where only the finer detrital materials could reach the site of deposition. The lack of iron also would indicate that they were deposited farther offshore. These beds are usually cemented with calcite and occasionally contain fossils; furthermore two thin coarse-textured sparry carbonate beds were found in this upper zone of the Rose Hill which is more calcareous and fossiliferous than the rest of the Rose Hill in an area extending from northern West Virginia into central Pennsylvania. These facts show that the upper Rose Hill was developing toward a carbonate-rich environment as a forerunner to much more calcareous Rochester and McKenzie Formations which overlie the Rose Hill Shale and Keefer Sandstone. Combining this information would imply that marine transgression occurred during deposition of the upper zone of the Rose Hill. Thus if it was deposited farther from the shore line, there may have been also a general decline in the rate of influx of clay from the source land; both factors would tend to favor a trend toward carbonate deposition.

Source area.—The Rose Hill gives direct

evidence of low-rank metamorphic and older sedimentary source areas, and indirect evidence of a minor plutonic igneous source (fig. 8). Much of the quartz silt has undulose extinction, and the common occurrence of chloritic and sericitic slate and low-rank phyllite fragments proves a partial metamorphic source. Their increasing abundance in the upper part of the formation together with the marked upward increase in abundance and grain size of discrete mica flakes indicates that this source became increasingly important as the source area was denuded or drainage changes occurred. In the Rose Hill, the metamorphic fragments are noticeably more chloritic and also slightly more coarsely crystalline than similar fragments in the Tuscarora, and the discrete micas are more abundant, of larger size, and contain a larger proportion of chlorite flakes than those in the lower formation. All of these facts indicate that the Rose Hill metamorphic source was of somewhat higher rank (nearly chlorite-grade) than that which provided detritus to the Tuscarora; perhaps it was more deeply eroded into more potently metamorphosed zones, or the place from which most Rose Hill detritus was derived was in a different location from that which produced the Tuscarora. Here a question arises, however. Is it possible that the increased amount of chlorite in discrete flakes and as a constituent of the slate and phyllite fragments is due to the high-iron environment in which much of the Rose Hill was deposited? In other words, were the metamorphic rock fragments originally sericitic but after deposition in the Rose Hill environment did they take up iron and become partially converted to chlorite, whereas other iron was precipitated as hematite cement, glauconite, or chlorite pellets? That this is *not* true is shown by these facts: (1) chlorite is abundant even in the non-red Rose Hill samples such as the gray silt-shales near the top of the section; (2) chlorite is very scarce even in the hematite-cemented lower red Tuscarora. Therefore the chlorite is believed to be legitimately detrital and is not a product of the iron-rich diagenetic environment.

The abundance of rounded silt grains of about .03 to .06 mm (in contrast to the angularity of the larger grains) points to a well-rounded siltstone as an important rock type in the source terrane. This sedimentary terrane contrasted greatly with the sedimentary terrane that supplied Tuscarora detritus. During deposition of the lower Tuscarora the source area supplied abundant chert in grains ranging from fine sand up to pebble size together with large amounts of subround to well-rounded medium to fine quartz sand grains. The chert source gradually became less important during the upper Tuscarora, virtually disappearing by the end of Tuscarora deposition; this decline continues through the Rose Hill, which has only a minute quantity of detrital chert, occurring only as silt-size grains in the basal part of the formation. It is interesting to note that chert is virtually lacking in the upper Rose Hill and all subsequent formations in the Silurian so that this source had permanently disappeared. The source of medium to fine, round quartz sand seems to have disappeared by the beginning of Rose Hill deposition also since nearly all the large quartz grains in the Rose Hill (which occur in very small amounts in the red siltstones) are subangular to angular and appear to be of plutonic igneous origin because they are single crystal units with straight extinction and contain a moderate to small quantity of vacuoles. Large grains of composite metaquartz, declining greatly in the upper Tuscarora, are similarly very scarce in the Rose Hill; apparently that source was also eliminated. Much of the quartz silt has undulose extinction even when embedded as single grains in the clay shales where they should have been protected from post-depositional straining on folding of the rocks. This may have come from the low-rank metamorphic source. The source of the straight-extinguishing quartz silt is not known; it could come from non-strained metamorphic sources or from the plutonic igneous source that apparently provided the large, angular straight-extinguishing grains and much of the upper Tuscarora sand.

The quartz slivers (probably discs in three dimensions) may have come from erosion of narrow quartz veinlets produced during folding and deformation of the source. Narrow quartz veinlets (0.1 to 0.3 mm thick) in the present-day Rose Hill samples are made up of elongate quartz grains, some almost fibroid, growing per-

pendicular to the walls of the veinlets with the slow ray parallel with the length. Thus these vein quartz grains have the same shape, optic orientation, and size as the detrital quartz slivers so abundant in the Rose Hill. Many of these elongate grains may have been produced by the metamorphic source (Bokman, 1952).

The chloritic and illitic-sericitic clay that forms the great bulk of the Rose Hill presumably was derived from erosion of older chloritic and illitic-sericitic shales (?), low-rank slates, and phyllites in the source area. Nearly all of this material was broken down by abrasion into individual flakes and went to form the clay shale beds, but very minor quantities survived as discrete metamorphic rock fragments. Diagenesis in the marine environment may have enhanced the crystallinity of illite and chlorite slightly, and the deep burial and intense folding may have also helped; but the clay mineralogy is believed here to be very largely a product of simple disaggregation of source rocks, with little chemical change. The reason for the over-all abundance of clay and silt in the Rose Hill is that the source area provided a great mass of very fine detritus from breakdown of slate and phyllite; metamorphism was not intense enough to build coarse quartz in any quantity, and older sandstones had been largely stripped off during Tuscarora deposition.

In summary, the source area probably consisted of a series of quartz siltstones and shales overlying a very extensive mass of weakly metamorphosed slate and low-rank phyllite, some of it of chloritic grade. This complex was shot through by thin quartz-filled veinlets and intruded by some outcrops of granitic rock.

Climate.—As is the case with the underlying Tuscarora and the overlying Keefer, small contributions of quartz from plutonic igneous sources are evident in the Rose Hill. Therefore the almost total lack of feldspar would imply a humid climate favorable for its decomposition. The large amounts of hematite associated with the Rose Hill here and elsewhere indicate a climate in which ferromagnesian minerals could be intensely weathered in the source and their residue carried to the sea.

The presence of kaolin in these relatively impervious marine shales can be explained only by the presence of kaolin in the source, because this mineral does not form in the marine environment. Kaolin was probably therefore developed in the soil during a warm and humid climate through weathering of the slates and phyllites. It is well known that kaolin is restricted to the nearshore zone in recent marine sediments (Weaver, 1958; Van Andel and Postma, 1954), and this harmonizes very well with the interpretation previously stated that the Rose Hill was deposited in a brackish-marine zone offshore from a series of deltas with extensive fresh-water influx due to heavy rainfall in the source.

KEEFER SANDSTONE

Regional Relations

The Keefer Sandstone, the middle formation of the Clinton Group, is a very thin but remarkably widespread sandstone with a sharply-defined top and bottom (figs. 4 and 7). Its maximum thickness in central Pennsylvania is approximately 60 ft, thinning to 25 ft in western Maryland and northeastern West Virginia. It thins gradually to the west (5 to 10 ft in the subsurface of western West Virginia) and the sands finally die out at about the Ohio-West Virginia line (Rittenhouse, 1949). As shown in figure 7, the Keefer Sandstone has two facies (Woodward, 1941; Swartz, and others, 1923). In central Pennsylvania and in the western Silurian outcrop belt of West Virginia (which skirts the west edge of the West Virginia panhandle) it is calcite-cemented, somewhat friable, contains some thin fossiliferous limestones and fossiliferous sandstones bearing brachiopods, crinoids, and trilobites, and often has a thin oolitic hematite bed at the top. It becomes still more limy to the north and west and is probably represented by the Irondequoit Limestone of western New York; it changes to shale passing northeast to Williamsport, Pennsylvania (Swartz, 1939). In southeastern Pennsylvania and in the eastern Silurian outcrop belt of West Virginia (which runs along the eastern edge of the West Virginia panhandle) it is very hard, heavily cemented by quartz, non-calcareous, unfossiliferous, and has no limestone or hematite beds. In these areas it is a consistent minor ridge-

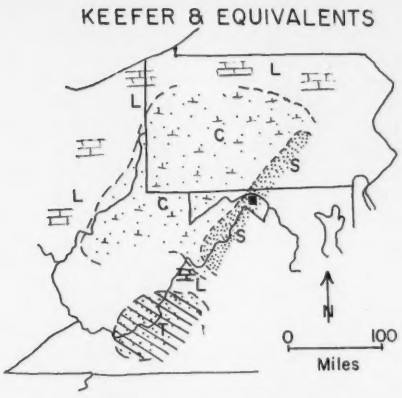

FIG. 7.—Areal variation of the Keefer Sandstone and equivalents. For further information see figure 4. The Keefer Sandstone is fossiliferous and cemented with calcite in the area labeled "C" but is unfossiliferous and cemented with quartz in the "S" area. It passes into shale and limestone in areas "L." The Keefer is represented by a very thick mass of Tuscarora-like sandstone ("T") in south-central Virginia. Location of the area studied is shown by the black square in panhandle West Virginia.

former, gray-white to brown, highly siliceous and resembles the Tuscarora except for slightly inferior hardness and finer grain and a more prevalent occurrence of *Scolithus* tubes (Woodward, 1941). In eastern Pennsylvania it possibly passes into the Shawangunk Conglomerate as do all underlying Silurian formations (Swartz and Swartz, 1931). According to Woodward (1941), in West Virginia it is coarser in the western outcrop belt with local zones of white quartz and jasper pebbles.

Just as it does in going towards northeastern Pennsylvania and western New York, the Keefer Sandstone rather rapidly thins and disappears when traced southward, being replaced by shale, fine-grained often cherty limestone, and crinoidal limestone near the Pendleton County, West Virginia—Highland County, Virginia line. But still farther south a very sudden change takes place in a belt from Rockbridge County, Virginia, through Mercer County, West Virginia (fig. 4). In a very short distance, an isolated body of hard, white Tuscarora-like sandstone 150 to 250 ft thick appears abruptly at about the Keefer horizon; it is not known whether this thick mass of sandstone is replacing the upper half of the Rose Hill (Butts, 1940) or post-Keefer beds perhaps as high as part of the Wills Creek (Woodward, 1941). This mass directly overlies the "Cacapon" red sandstone facies of the Rose Hill and appears to be separated laterally from the type Keefer by a short limestone-shale area; it may have developed independently from the same local uplift that has been postulated as the source of the "Cacapon" Facies.

The section studied here is in the eastern outcrop belt; thus it is heavily cemented with quartz and contains virtually no carbonate or fossils.

The prevalence of quartz cement near the source area in the southeast and the abundance of calcite cement in the more offshore parts of the Keefer may be a more widespread phenomenon. Similar relationships have recently been observed in the Pennsylvanian Tensleep Sandstone of Wyoming by Todd (1959), who found that quartz cement became relatively more abundant compared to carbonate cement as the source area was approached toward the north or northeast. The same changes have also been found by L. T. Rogers (M.A. thesis in preparation, The University of Texas) in the Strawn (Pennsylvanian) Sandstones of north central Texas, which are calcite-cemented in the more offshore zones and have more quartz cement toward the source area in the southeast. Presumably silica derived from weathering in the source area was precipitated fairly rapidly upon encountering sea water, and calcite cement dominated in the areas beyond.

Local Character

A complete section of the Keefer Sandstone is exposed along a cut of the Baltimore and Ohio Railroad approximately 1 mile southwest of the former station at Sir Johns Run, approximately 2 miles northwest of Berkeley Springs, Bath Township, northern Morgan County, West Virginia. The Keefer here is 20 ft thick, and the contacts with the adjacent shale formations are quite sharp. The section was sampled at 2 ft intervals, and extra samples were taken of unusual lithic types.

The Keefer at Sir Johns Run consists of two very distinct divisions. The lower 8 ft of section is a very heterogeneous, dark colored and irregularly bedded mixture of rocks of every textural grade between clay shale and granular coarse sandstone with some grains as large as 4 mm. The sediments are subangular, poorly sorted, and texturally immature and range between orthoquartzite and subgraywacke in composition. The upper 12 ft of the formation offers an extreme contrast, consisting of pale brown, exceedingly uniform, homogeneous, massively bedded, superbly well-sorted and highly rounded fine-grained sandstone. These beds are texturally supermature, have an extreme orthoquartzite composition, and are heavily cemented with quartz. In other localities in Morgan County this simple division does not always hold true although the dark shaly beds commonly occur in the lower part and the well-sorted fine sandstone beds in the upper part.

In the Keefer as a whole all the white to light gray or pale brownish rocks are pure orthoquartzites. The darker gray rocks are all either subgraywackes or else immature orthoquartzites with a clay matrix. Thus, in general, the darker gray rocks indicate a relatively low-energy depositional environment, one in which the soft metamorphic rock fragments were preserved from destruction and/or in which clay matrix was allowed to accumulate.

Lower zone.—The lower zone shows extreme textural variation, but is mostly sandstone with a small proportion of shale interbeds. The sandstones range from N6 or N7 (medium light gray to light gray) to N3 or N4 (medium dark gray to dark gray) depending on the proportion of dark clay matrix; all the dark ones are immature. The shales are N3 (dark gray). Upon weathering these rocks become covered with a thin brownish limonite film because of the oxidation of pyrite.

The sandstone beds range from 1 to 3 in thick, and the bedding planes are hummocky and irregular. Furthermore, many of the sandstones contain black, undulose clayey seams 0.1 to 1 mm thick spaced 2 to 5 mm apart. They are very hard because of the clay matrix and quartz cement.

The sandstones in this section contain a gamut of grains from clay and silt through all the sand grades up to granules as large as 3–4 mm (plate 6, D). Each bed has a different textural name representative of the group being clayey coarse to very fine sandstone and granular very coarse to fine sandstone. Even the coarse sandstones may have a clay matrix. These samples are all bimodal or polymodal. A clay or silt matrix is common; a second common mode ranges between 2ϕ and 4ϕ (.25 to .06 mm), usually about 2.5ϕ (.18 mm); and a third mode varies between -1ϕ (2 mm) and 1.3ϕ (.40 mm) with an average of about 0.5ϕ (.70 mm). The coarsest mode and the finer sand mode may vary in all proportions, and the clay or silt matrix may or may not be present in any of these. In the coarser beds the largest grains found are approximately 2 to 4 mm. These grains are composed of clear to somewhat milky quartz.

The great variation in grain size is accompanied by similar variation in textural maturity. Some beds are immature, containing abundant clay matrix and poorly sorted quartz grains; even large quartz grains in these samples are subangular. Others lack clay but are poorly sorted (submature) and some beds are quite well sorted and contain well rounded quartz grains, hence are supermature (plate 6, E). These marked variations take place within stratigraphic intervals of 3 ft or less.

Most of the gray sandstones show disturbed bedding and swirled structures evidenced by the orientation of the grains probably caused by burrowing organisms.

The dark gray shale interbeds are a peculiar rock type. They consist of imperfectly fissile, hard dark gray clay shale, rather rich in fine mica and illite. But within this clay occur undulose or curdled 1 to 3 mm laminae, irregular nests, swirled masses, and lumps of light gray silt to subangular fine sand (plate 6, F). These masses are sharply differentiated from the surrounding dark gray clay, averaging about 2 to 8 mm across and form 15 to 20 percent of a given specimen (plate 7, A). Each lump consists of well-sorted coarser grains (coarse silt to fine sand) in a clay matrix. Within an individual lump all quartz grains are about the same size, but there is great variation in sand grain size comparing the lumps with

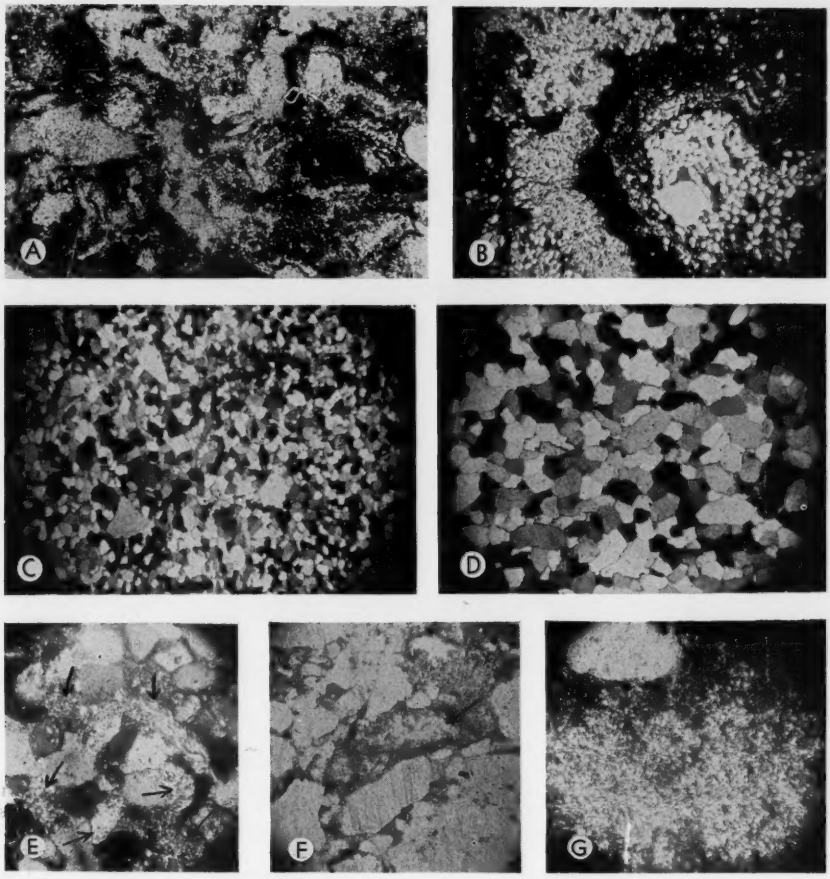

PLATE 7.—Photomicrographs of Keefer Sandstone.

A. Silt lumps or "mottles" in black shale, lower Keefer. Note variation in grain size of the quartz particles within lumps. These are probably caused by burrowing organisms. Ordinary light, ×3.

B. Same as A, showing detail in three lumps. Grains in each lump are fairly well sorted. Ordinary light, ×13.

C. Upper Keefer Sandstone, a supermature orthoquartzite. This rock is extremely well sorted (except for a minor coarse mode), and grains are very well rounded. Grains are virtually all common quartz, and the rock is cemented tightly with interlocking quartz overgrowths. Crossed nicols, ×11.

D. Same as C, showing extreme purity of this orthoquartzite. Crossed nicols, × 32.

E. Immature subgraywacke sandstone in the lower Keefer, with abundant metamorphic rock fragments (arrows) and micas. Contrast the mineralogy of this bed with that of D, only 3 ft stratigraphically above; the difference is believed due to intense abrasion in the beach-dune depositional environment of the rock in D. Crossed nicols, ×110.

F. An immature orthoquartzite in the lower Keefer. The gray clay matrix is replaced in one area (arrow) by well-crystallized illite (clear, in middle of clay patch). Ordinary light, ×38.

G. Same as F, detail of illitic replacement area (brightly birefringent) in middle of gray clay patch. Crossed nicols, ×130.

each other (plate 7, B). These features have been termed "mottles" by Moore and Scruton (1957) who have found that these structures in recent sediments are caused by the activities of organisms. Thus they are akin to the swirled structures found also in the sandstones of the lower zone.

Upper zone.—The upper 12 ft of the Sir Johns Run section is exceedingly uniform in all characteristics. Beds average 5Y to 10YR7/1 (yellowish-gray to light brownish-gray), weather brownish, and occur in beds 6 in to 2 ft thick. These beds are so strikingly uniform in grain size that they are devoid of any visible lamination or cross-bedding. Beds are very hard because of abundant quartz cement, break across the grains with a smooth conchoidal fracture, and are true sedimentary quartzites. All beds are classified as very well sorted fine sand with the mode at 2.4 to 2.9ϕ (.13 to .19 mm). A minor very well sorted coarser mode at .35 to .60 mm (medium-coarse sand) makes up 1 to 10 percent of these sandstones (plate 7, C), and the coarsest grains in the slides are 0.3 to 1.0 mm. The standard deviation is 0.25 to 0.35ϕ (very well sorted). All quartz grains are well rounded, hence the upper zone is texturally supermature (plate 7, D).

Occasional vertical borings (*Scolithus*) about 3 to 6 ins long and $\frac{1}{8}$- to $\frac{1}{4}$-inch in diameter are present. These are filled with sand of the same grain size as the matrix but are slightly darker. In the bulk of the rock, grains are well oriented parallel with the bedding.

In a section measured on the railroad cut $1\frac{1}{2}$ miles east of the town of Great Cacapon, Morgan County (about 2 miles south of the Sir Johns Run section) a 6 in bed of granular very coarse sandstone appears at the very top of the section, and there is also one thin bed of dark gray clayey very fine sandstone; otherwise the upper Keefer here is like the upper zone at Sir Johns Run.

Descriptive mineralogy.—The two zones show as sharp a contrast in mineralogy as they do in texture. The lower zone contains significant amounts of mica and metamorphic rock fragments, and therefore it is classed as borderline between orthoquartzite and subgraywacke (plate 7, E). Terrigenous constituents in the upper zone consist entirely of quartz except for slight traces of tourmaline and zircon; a typical sample is classed as very well sorted fine sandstone: highly siliceous supermature orthoquartzite (plate 7, C, D).

Quartz in the Keefer, even in the lower micaceous zone, is almost entirely common quartz, that is, single crystal units with straight to slightly undulose extinction. Less than 5 percent of the grains have abundant vacuoles (hydrothermal), and composite quartz grains are very rare. Even the large grains (0.5 to 2.0 mm) are almost entirely common quartz and are no different in properties from the smaller grains (plate 6, D, E). A few grains have inclusions of apatite or rutile needles, but microlites are very scarce. About one-fourth of the grains have undulose extinction, but this is not significant because the rocks have been strongly folded and are transected by oriented bubble trains which formed during folding. The more immature beds in the lower zone contain quartz grains that are dominantly subangular (plate 6, D), but there are well-rounded grains admixed with them in significant quantities (about one-fourth of the total). Both the larger and smaller quartz grains seem to have about the same proportion of subangular and round grains. The more mature beds in the lower zone, as well as the supermature beds in the upper zone, consist entirely of rounded grains (plate 6, E; plate 7, D). The subangular grains of all sizes show the same overwhelming predominance of common quartz as do the rounded grains, showing that the lack of composite quartz, metaquartzite, and vein quartz is caused by their rarity in the source area; it is not due to selective abrasional destruction of the composite grains.

Almost all the quartz grains (except in the clayey sands) have large overgrowths which fill up the entire pore space of the rock.

Only one small subround grain of chert and three grains of sericitized, subround orthoclase were seen in the Keefer slides, and these all occurred in one specimen 6 ft above the base.

Large micas average 0.1 to 0.5 percent in the lower part of the section and hence are quite conspicuous. Muscovite forms about three-fourths of the mica with pale brown

biotite and pale green chlorite making up the remainder. Biotite is most common in the same specimen that contained the orthoclase. Mica occurs in large, thick books averaging .10 to .30 mm long and thus is considerably coarser here than in the underlying Rose Hill.

Metamorphic rock fragments are entirely lacking in the upper Keefer, but in the lower zone range from 0 to 20 percent, averaging about 5 percent (plate 7, E). Grains are round, elongate, and .05 to .2 mm in diameter; thus none occur in the coarser mode. They range from slate to low-rank schist, most of them showing good foliation and probably classifiable as high-rank phyllite. They contain more mica than quartz, and the constituent mica shreds are .01 to .05 (rarely .10) mm long, consisting very largely of sericite with a little chlorite. The recrystallized quartz in these fragments averages .005 to .02 mm in size. There are a few grains of stretched metaquartzite, lacking micas.

One sample from the Great Cacapon section contained a few penecontemporaneous siltstone fragments.

Tourmaline grains average .10 mm. They are well-rounded in the sandstones that consist of well-rounded quartz and subangular to subround where the associated quartz is also subangular to subround; these variations in roundness occur in beds that are only a few feet apart stratigraphically. This shows that the tourmaline was being rounded during Keefer deposition, probably at about the same rate as the quartz grains. Green and brown grains are most common, but there are some colorless, orange, and gray-blue ones. Many tourmaline grains are fractured by cross-cutting tectonic quartz veinlets. In some samples, the tourmalines possess authigenic tourmaline overgrowths. These are idiomorphic against quartz overgrowths but abruptly stop at the detrital quartz grain margin. This indicates that authigenic tourmaline grew prior to quartz cementation and did not grow by replacing quartz but formed a directly precipitated pore space filling. Zircon is also present in subround to subangular colorless grains, and leucoxene is also common.

The clay matrix in the immature sandstones of the lower zone appears to be mostly illite, and it is rich in larger flakes of mica. The material is black or dark brown, apparently high in organic matter.

One specimen from the Great Cacapon section showed some dolomitized ghosts of probable fossils together with 1 to 3 mm masses of isotropic, rich brown collophane with included silt grains, probably large coprolites. Where these coprolites are crossed by tectonic fractures, colorless hypidiomorphic apatite crystals developed authigenically.

Quartz overgrowths occur in all slides. In the clayey sandstones of the lower zone only small amounts are present, but in the better sorted ones quartz is very abundant (up to 10 to 15 percent of the rock) and completely fills all pore space in an interlocking mosaic. Overgrowths formed prior to regional folding, because tectonic bubble trains pass through them. Other than this, their time of origin is uncertain.

Ankeritic dolomite, some of it limonite-stained, appears in a few slides in the lower zone as .1 to .2 mm crystals. These are always idiomorphic against authigenic quartz, but normally abut against detrital quartz with the exception of one slide wherein it replaces the edges of a few detrital quartz grains. Thus dolomite, like tourmaline, formed by direct precipitation into pore space prior to crystallization of quartz overgrowths. In the clayey rocks traces of pyrite occur in the organic-rich clay laminae, and in non-clayey rocks there are limonite grains which were probably derived from pyrite.

Authigenic chlorite occurs in subequant to attenuated .2 to .5 mm patches in many slides in the lower part of the section. The color ranges from pale green to khaki to almost colorless, and the interference color is anomalous blue gray to pale gray; the mineral occurs as randomly oriented shreds and scales 2 to 25 μ in length. Some of these patches also contain books of colorless, water-clear, yellow-birefringent illite. Sometimes illite occupies the center of the patch with chlorite surrounding it; at other times the reverse is true. These micaceous minerals are usually idiomorphic against quartz overgrowths or encrust the detrital grains preventing the development of the overgrowth. Some thin coatings of micaceous mineral occur between detrital quartz grains and

their overgrowths. These facts indicate that the micas formed first. Because of the large size of the patches, it is probable that the mica is replacing some other substance. In some slides it is apparent that detrital clay stringers and patches have recrystallized to these coarser, authigenic micas because detrital clay on the outside edge of a patch changes gradually into authigenic, well crystallized mica on the inside of the patch (plate 7, F, G).

After deposition, some suturing and solution of quartz has taken place under pressure. Some detrital, originally rounded grains interpenetrate along sinuous to almost stylolithic contacts (plate 6, E). Suturing is common (but not universal) in the clayey sands but almost non-existent in the clean ones. One slide is a striking example of this. In clean parts of the slide, suturing is virtually absent, but extensive quartz grain suturing has occurred along clay stringers in the very same slide. This confirms the view of Heald (1955) and Thomson (1959) that clay films promote solution of quartz under pressure.

Depositional environment.—The lower zone of the Keefer, with its closely interbedded rock types of great variability in grain size, textural maturity, and detrital mineralogy, must represent deposition in two or more environments that were adjacent and rapidly fluctuating. Furthermore, one of these environments must have been a high-energy one to cause the sporadic good sorting, rounding of quartz, and elimination of metamorphic rock fragments by abrasion, and the other environment must have been a low-energy one to produce the poorly sorted, angular, clay-rich sandstones and black shales that are rich in metamorphic material. These two environments fluctuated rapidly in order to produce interbedding of their characteristic rock types on a scale of a few inches. Further evidence of origin is the abundant borings and organic-caused "mottles" in the black shales and clayey sandstones and the presence of pyrite which indicates reducing conditions. The picture that seems to fit these facts best is a depositional locus straddling an area of a barrier bar (or beach) and a protected semi-stagnant lagoon. Here is an ideal situation with an extreme high-energy environment immediately adjacent to an extreme low energy one, and slight shore-line changes can easily accomplish a mixture of the two phases. Some of the dominantly angular sediments swept down from the source area were deposited in the beach environment where intense abrasion took place sufficient to cause excellent rounding and sorting of quartz during Keefer deposition, and complete removal of rock fragments and clay produced light-colored supermature orthoquartzites. Currents were strong enough to transport grains as large as coarse sand or granules. Other sediments accumulated perhaps on the back side of the barrier or on the fringe of the mud zone, where there was not so much wave action; these retained their angular and subangular quartz even in grains as large as 1 to 2 mm. Lack of abrasion allowed preservation of metamorphic rock fragments and did not winnow out the micas; thus most of these rocks retained their source area composition as subgraywackes. The energy of the environment was low enough that the black clay and mud of the lagoon did not get washed out of the sands; thus these sediments are immature and dark colored. At times sediments that had once been on the high-energy beach were washed back into the lagoon, producing textural inversions such as well-rounded coarse-grained orthoquartzites that are dark colored and have an abundant clay matrix. All these fringe-line sediments, deposited on the edge of the lagoon just in back of the barrier, were extensively burrowed by organisms. The relatively minor black shales with silt mottles and sandy lumps were apparently deposited farther back in the stagnant, reducing lagoon where only minor admixtures of sand were carried in from the adjacent barrier bar. Moore and Scruton (1957) have shown that such mottles form characteristically near the boundary line between two environments.

It cannot be too greatly emphasized how markedly the lower Keefer varies from bed to bed as a function of the environment of deposition. Quartz roundness and metamorphic rock fragment (MRF) content vary in close harmony with each other. A bed at 7 ft contains 20 percent MRFs and micas and consists of angular quartz grains while

beds at 3 ft and 10 ft contain rounded quartz grains of the same size and have almost no MRFs. This rounding of quartz and elimination of metamorphic rock fragments by abrasion took place during Keefer deposition, sometimes in such short-lived intervals that only a couple of feet of rock were laid down. This considerably disparages the ideas (1) that quartz grains to become rounded must be recycled many times through successive geologic formations or (2) that rounding requires prolonged tectonic stability. Rounding can be accomplished rapidly (in stratigraphic terms) in a single cycle of deposition if the environment is potent enough.

The upper zone of the Keefer appears to be a very uniform supermature beach or barrier bar sand of extreme orthoquartzite composition. Either the environment had migrated to place the future Sir Johns Run section well into the pure barrier bar, or the lagoon may have disappeared. Presence of the clayey, poorly sorted sand characteristic of the basal zone as tongues in the upper zone in other sections (for example, Great Cacapon) shows that the same conditions prevailed elswhere in the area. Why is the upper zone, which is a pure fine sand, so much finer than the lower zone of immature clayey sediments which have everything up to granule size? This could be explained as follows: the hypothetical environment contained a semistagnant lagoon with black mud, lagoon beach, probably a belt of dunes down the interior of the barrier bar, an open-coast beach, and finally the open marine area. The mixture of coarse sand and clay is probably characteristic of the beach-lagoon or beach-marine borderline where waves of high energy capable of carrying coarse sand and granules were present; yet the nearness of the lagoon allowed frequent mixing with clay to take place. The well-sorted, extrmely uniform sands of the upper zone may well represent sands accumulated in the dune belt near the middle of the hypothetical barrier bar. The winds would be ideal for producing excellent sorting and rounding. The lack of clay and coarse grains could be explained by having the coarser grains lag behind on the beach and only the fine sand carried to the middle of the island and heaped up as dunes. Yet, these fine sands have vertical borings, ascribed to *Scolithus*, allegedly a marine worm. Therefore, on final deposition a marine transgression must have occurred, submerging the dune sands, swashing them around to destroy cross-bedding, and allowing them to become riddled with *Scolithus* tubes.

Mainland beach versus barrier bar origin.—In the discussion above it has been assumed that the Keefer formed on a barrier bar, separating a lagoon or bay from the open sea. Why could not it have also formed as a beach fronting the mainland? There is no direct way to decide between these two alternatives, because research in recent sediments has not yet enabled us to distinguish a barrier bar from a mainland beach by physical properties of the sediments themselves. By consideration of the adjoining formations, however, it is possible to gain some insight into this problem. The underlying Rose Hill Shale appears to be a normal open-water marine sediment because it is a thick section of light olive-green clay-shale with a few carbonate beds and calcite-cemented siltstones in its upper portion. The overlying Rochester Shale, on the contrary, consists largely of very black, fissile clay shale with some intercalated calcirudite to calcilitute limestones. This formation is only 33 ft thick, and its evidence of shallow-water yet stagnant conditions would seem to fit best with a lagoon environment of poor circulation. The small thicknesses of the Keefer and Rochester would fit well with the small geographic extent of a barrier bar-lagoon couplet at any one time.

If the whole sequence is considered as a transgressive environment then one could interpret the Rose Hill as open-water marine, passing through the Keefer (barrier bar) into the Rochester stagnant lagoonal sediments. Although it may not be correct, it seems to yield a more harmonius picture than a sequence from open marine through mainland beach then reverting to lagoon.

Source area.—Because of the presence of subangular sands that have suffered little abrasion in the Keefer depositional environment, one can postulate rather well about source area. Softer constituents have not been eliminated by abrasion and the detritus reflects fairly closely the composition of the source. Lower-rank metamorphic rocks,

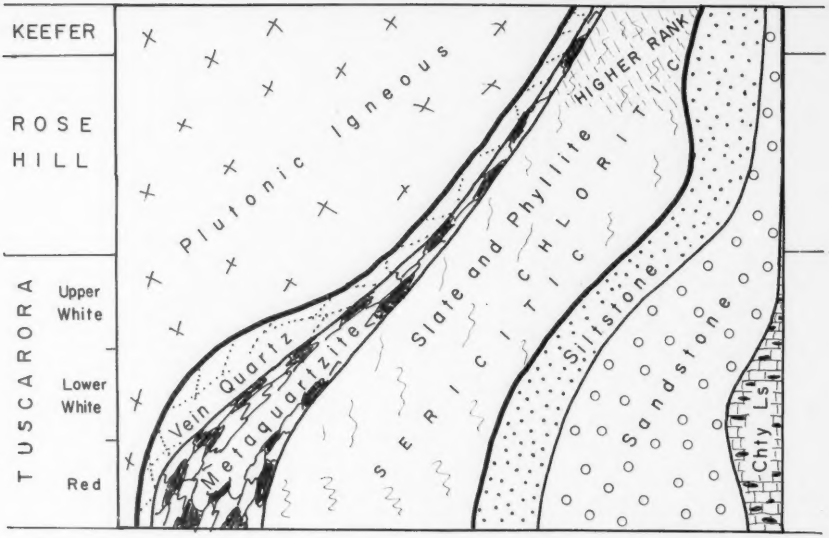

FIG. 8.—Proposed changes in source area lithology during deposition of Tuscarora, Rose Hill, and Keefer Formations. This diagram is a careful interpretation and shows what proportion of the detritus came from each source and how these proportions changed through time. A sedimentary cover was stripped off first, sandstone and cherty limestone all but disappearing by early Rose Hill deposition. Low-rank metamorphics were important throughout but diminished slightly upward; their rank increased, however, as shown by the coarsening of the micas; chlorite also became more prevalent upward. Metaquartzite decreased sharply upward, and vein quartz culminated in the lower white Tuscarora (partly a function of size since this unit has most pebbles). Plutonic igneous sources were minor at first but rapidly became of great importance as the mantling sediment and metamorphics were stripped off by erosion.

sericitic slates and phyllites, were important contributors as they were throughout the Tuscarora and Rose Hill (although they were removed by abrasion in the upper Tuscarora and upper Keefer). This is shown not only by the presence of discrete slate and phyllite fragments but also by the abundance of individual mica flakes. The rank of the metamorphic rock fragments was somewhat higher in the upper Rose Hill and lower Keefer than in preceding beds, as shown by the coarser crystallinity of the micas in the rock fragments and also by the larger size of the free mica flakes (fig. 8).

Apparently the source of metaquartzite and composite quartz grains, which was abundant during lower two-thirds of the Tuscarora and was gradually disappearing through upper Tuscarora and Rose Hill, had all but completely disappeared during Keefer deposition. The same is true for vein quartz (semi-composite and highly bubbly grains). Even in the coarser and more angular Keefer sands, there is very little vein quartz, composite quartz, and metaquartzite in contrast to the Tuscarora where such types are very abundant; in fact the Keefer has almost entirely common quartz. The lack of these quartz varieties hence is not caused by their selective removal by abrasion but is due to a change in the source area. Apparently, most of the quartz of the Keefer came from plutonic sources, probably granitic judging by the abundance of straight-extinction grains.

In the dominantly subangular sands, there are admixed a small proportion of subround to well-rounded grains; these rounded grains range from coarse sand to very fine sand size. These grains have been inherited, either from previous sedimentary rocks or else they may represent stray grains from the high-energy Keefer environments, blown or washed into the low-energy Keefer

environments. It is not possible to decide between these two potential sources but the latter view is favored. Virtual lack of chert in Keefer, Rose Hill, and upper Tuscarora points to disappearance of the chert-bearing carbonates in the source area.

As an illustration of the possible control of source area lithology upon texture, the fact that the Keefer at Sir Johns Run has no grains larger than 4 mm may be the result of a granitic plus slate-phyllite source area with few veins or coarse metaquartzites. Few quartz grains in granite are larger than 2 mm (Dake, 1921), and slate or phyllite has not reached a high enough stage of metamorphism to provide coarse quartz. Therefore, quartz grains of large size were not available in the source during Keefer deposition. The Tuscarora may be coarser because its source area had abundant hydrothermal veins and coarse metaquartzites in it which provide quartz pebbles of large size; energy levels, stream gradients, and other factors may have been equivalent for both the Tuscarora and Keefer, but the disappearance of vein quartz and metaquartzite from the source may be the most important factor in the changes in grain size.

Climate.—Evidence on the climate during Keefer deposition is weak. If, as it appears, most of the Keefer quartz came from granitic rocks, the lack of feldspar (even in the more angular sands) would imply a humid, warm climate in the source.

SUMMARY

By integrating conclusions reached about depositional environment and source into a coherent picture for all three formations one can visualize a gradual and fluctuating transgression of the sea as the source area declined in elevation and relief after the close of the Taconic Orogeny. The source continued, however, to shed enough detritus throughout this time so that the Tuscarora, Rose Hill and Keefer consist of more than 99.9 percent terrigenous rocks in the area studied. During this long period of erosion the sedimentary blanket on the source area was largely removed and plutonic igneous areas became exposed; low-rank metamorphic sources were important during the entire interval, but contributions decreased somewhat in the later stages of deposition.

The subaerial delta of the Juniata slowly became submerged and marine or brackish waters flooded up the drowning river mouths to produce estuaries. The red Tuscarora was deposited in these shallow, oxidizing waters; current direction was consistently northwest, but the amount of abrasion was insufficient to remove soft metamorphic rock fragments or to produce any quartz rounding, and iron was precipitated upon encountering saline water. In this transitional brackish environment lived the organism that produced the *Arthrophycus* imprints. Sluggish stream gradients over a broad, flat coastal plain prevented transport of any gravel into the area. Next, however, a significant uplift of the source area took place; streams quickened as gradients increased, and floods of vein quartz pebbles were swept in to begin deposition of the white Tuscarora. These fluvial gravels were attacked by surf action along the coast and a swath of gravelly beaches developed. Most of the metamorphic rock fragments were ground up by wave action and transported seaward as illite or chlorite particles and quartz silt; some rounding of sand-size quartz was accomplished but not to an extreme degree because the influx of sand and gravel was too rapid. As highlands resulting from this uplift wore down, stream gradients decreased and gravel came in declining quantities with sporadic minor renewals. The beaches had ample time to work on the grains. As a result of the slowed influx of detritus, excellent rounding and sorting of the medium sand was accomplished and the more randomly cross-bedded supermature orthoquartzites of the upper part of the white Tuscarora were produced. All metamorphic rock fragments were eliminated. Debris from disaggregated slate and phyllite fragments moved seaward to accumulate in the marine environment as the thick beds of illite-chlorite shale. These deposits are now designated as the Rose Hill. During Rose Hill deposition, the source area was continually lowered; streams still brought in large quantities of clay and some silt, but gradients were by now so low that virtually no sand reached the depositional area and the shoreline had retreated a considerable distance east. Seasonal rainfall in the source area may have periodically freshened the water, inhibiting life so that fossils are quantitatively rare in the shale beds. Peri-

odic incursions of silt were laid down in beds 1 to 3 in thick, perhaps caused by occasional storms or temporary shifts in current threads.

During middle Rose Hill deposition a minor local shoaling of water took place and the red very fine sandstones were deposited at intervals in a more turbulent, oxidizing brackish-marine environment probably off the mouths of large rivers where iron was being carried down to the sea. There were many such shoalings occurring at different places and at different times during Rose Hill deposition (Castanea, "Cacapon," Cresaptown—all iron sandstones). Influx of terrigenous clay and silt slowed somewhat during upper Rose Hill deposition, and the first carbonate beds were laid down. However, these were never more than a few inches thick and occur very rarely in the section.

It is probable that a renewed, but very minor, uplift of the source produced the sands of the Keefer. In northeast West Virginia, the Keefer appears to have accumulated as a barrier bar separating stagnant, black lagoonal clays (now the Rochester black shale) from the more open marine olive green clays of the Rose Hill. The lower Keefer is made up of sediments from a transtional environment where barrier beach sands were mixed with lagoonal clays which were thoroughly churned by organic burrowing. At sporadic locations or times, some of the Keefer sands were exposed to sufficient surf action to become well-rounded orthoquartzites while at other places or times, sands from the same sources, through lack of modifying action, retained their original angularity and metamorphic rock fragment content and are thus immature subgraywackes. In the upper Keefer the barrier bar became fully developed with extensive dune tracts; these produced the superbly sorted and rounded, remarkably uniform, almost completely pure quartz sands of the upper Keefer. As the Keefer barrier bar moved westward, it was succeeded in the section by the Rochester lagoonal shales.

Turning to the character of the source area, only that part of it can be discussed which was furnishing detritus to the area that is now in northeastern West Virginia. Studies by Tuttle (1940) and Krynine (1946) show a similar source terrane for the rocks deposited in central Pennsylvania, but nothing is known of the paleogeology of the source farther south because thin-section petrography has not been done.

Probable source area evolution is shown in figure 8. Evidence shows that the source during Juniata and red Tuscarora deposition consisted of extensive slate and phyllite with some metaquartzite and granite overlain by large areas of older medium to fine-grained supermature orthoquartzite sandstone, siltstone, and some chert-bearing limestone or bedded chert. During deposition of the lower part of the white Tuscarora, the chert source was gradually and almost completely eliminated and a large amount of vein quartz began to be contributed as pebbles. By upper white Tuscarora deposition, even the vein quartz, metaquartzite, and sandstone sources had nearly disappeared, and granite was contributing abundantly. During Rose Hill deposition the source contained much granite, slate, and phyllite, but some orthoquartzite siltstone was still present. Weathering of the granitic source, first uncovered extensively during upper Tuscarora deposition, may be responsible for the high iron content of the Clinton (Miller and Folk, 1955). The granite provided not only magnetite, but also hornblende, biotite, and other mafic minerals which, when weathered, would release iron to be carried to sea and precipitate as hematite on contact with marine water.

In the upper Rose Hill metamorphics of somewhat higher rank (more chloritic; coarse mica and more coarsely crystalline phyllite) were uncovered, and these continued to supply the overlying Keefer. The Keefer source, in addition to these slates and phyllites, consisted of much granite and a little quartz siltstone but no chert. There was only a small amount of vein quartz and little or no sandstone (fig. 8). In both the Keefer and Rose Hill the clay fraction came largely from disaggregation of slate-phyllite while the sand was mostly of granitic origin. Silt was obtained from both metamorphic and older sedimentary sources

In none of these earlier Silurian formations is there any evidence of volcanic contributions. Thus the source was *not* a conventional volcanic island arc; rather it was a non-volcanic welt with a sequence of un-

metamorphosed sandstones, siltstones, and cherty limestones overlying low-rank metamorphics, hydrothermal veins, and plutonic igneous rocks. There is no evidence from this study to show whether the source was (a) a chain of disconnected non-volcanic islands, (b) a continuous narrow strip of non-volcanic land with culminations at various places, or (c) a fairly broad landmass (for example, Appalachia). However, the large volume of sediment and the consistent cross-bedding directions in the northern Appalachians lead this writer to favor possibility (b).

Evidence of a granitic source in abundance for the upper Tuscarora, Rose Hill, and Keefer sandstones combined with the almost total lack of feldspar favors a warm humid climate. Furthermore, the great quantity of iron minerals (hematite, ankerite, chlorite) formed in the near-shore environments of the Rose Hill and red Tuscarora would necessitate extensive weathering to produce the required amount of dissolved iron. Kaolinitic clays of the Rose Hill also must have developed in the soils of a source area that had heavy rainfall and warm temperatures to produce thorough leaching. Relief of the source was probably great enough to provide a mixture of heavily weathered soil clays with relatively fresh unaltered fragments of slate and phyllite from canyon sides (Weaver, 1958).

Petrologic conclusions.—Below are listed petrologic conclusions that may be of more than local importance:

1.—Stratigraphic changes in mineral composition for the formations studied are chiefly a function of source area and arise from the progressive denudation of a source land with the classical sequence of older sediments on top, then low-rank metamorphics, and finally plutonic igneous rocks. There is no evidence for any volcanic activity in the source welt or borderland ("Appalachia") raised by the Taconic orogeny. This borderland is regarded as a "line" source with culminations, rather than a "point" source. Tectonism was most intense and relief was greatest during deposition of the lower part of the white Tuscarora, and the welt gradually wore down and relative stability prevailed later on.

2.—Evidence from (a) degree of rounding accomplished on quartz grains in the final depositional site, (b) amount of destruction of soft metamorphic rock fragments, (c) degree of winnowing of silt and clay matrix, (d) sorting of the sand fraction considered independently of the gravel or mud fraction, and (e) randomness and dip of cross-bedding, all integrates to indicate a low-energy estuarine environment of little sorting and no abrasion for the lower (red) Tuscarora, a high-energy beach environment overwhelmed by rapid influx of detritus for the lower part of the white Tuscarora, and a constant high-energy beach environment with slow influx of detritus for the upper part of the white Tuscarora. Points (a) through (d) also indicate relative sorting and abrasional potency of the Keefer lagoon-beach-dune environments while modal size and sorting seem to be the main reflection of the physical characteristics of the environment during Rose Hill neritic deposition. Coarser sediments seem to be the deposits of more shallow marine waters of the Rose Hill.

3.—High rounding of quartz grains of fine to medium sand size was efficiently accomplished by Silurian beach environments in one cycle of deposition. This is demonstrable in the lower part of the white Tuscarora and especially in the Keefer where angular sands may alternate with well-rounded sands in beds less than 5 ft apart, all sands coming from the same general source. Allowance must, of course, be made for the inherited well-rounded grains which are present in all sands; thus the mean roundness of the sample is an inadequate measure, and one must examine the degree of rounding of the most angular grain population present.

4.—Metamorphic rock fragments (slate, phyllite) are very valuable environmental indicators. They are present in great abundance in the low-energy deltaic, lagoonal, fluvial, and estuarine sands (Juniata, red Tuscarora, parts of Keefer) and completely removed in a stabilized beach environment (white Tuscarora, parts of Keefer). They are present in moderate amounts in Rose Hill sediments, because most of these fragments were removed as the sediments passed through a beach phase in route to final deposition in the meritic zone.

5.—Chert grains of sand size wear out selectively compared with quartz in the beach environment, but the difference is

slight. There is no evidence whether or not metaquartzite (composite quartz) wears at the same rate or faster than single quartz grains. Because composite metaquartzite is very rare even in the coarse and angular sands in the upper part of the formations studied, its disappearance is caused by elimination in the source rather than by heightened abrasion and selective destruction.

6.—The source area during deposition of the Tuscarora, Rose Hill, and Keefer Formations provided four chief modal sizes of detritus, each one of distinct geologic origin, which combined in various proportions to form the various textures found in these beds. These modes are (1) a small pebble-to-granule mode (about 2 to 10 mm) composed largely of vein quartz; (2) a medium sand mode (.25 to .40 mm) composed mostly of quartz derived from plutonic igneous rocks or else fro n older sandstones that were in turn derived from granitic rocks; (3) a very fine sand-coarse silt mode (.03 to .10 mm) of problematic origin but much of it probably from disaggregation of fine-grained low-rank metamorphic rocks like silty slate or weakly recrystallized phyllite, some of it from older unmetamorphosed siltstone; and (4) a clay mode, mostly illite, from disaggregation of clay slates and the fine-grained hydromica portion of phyllite. There is a lack of sizes intermediate between these modes. The Juniata and red Tuscarora contain mostly mode (2) with small amounts of (3) and (4); the white Tuscarora is mostly (2) with a small amount of (1); the Rose Hill is mostly (4) with considerable (3) and traces of (2); and the Keefer is mostly (2) and (4) with smaller amounts of gradational sizes.

The bimodal character of the gravelly sandstones in the Tuscarora is caused by the different origin of the two size modes; the bimodal character of the Rose Hill (well-sorted siltstones plus well-sorted, nearly pure clay shales) is also caused by the difference in genesis of these two distinct size modes.

The similarity in modal sand grain size between the very different environments of the red and white Tuscarora indicates that source area exerts the dominant control on sand size properties. The environment of deposition was effective in winnowing out clay and silt and in rounding the quartz sand grains but did not effectively change the modal grain size of the quartz.

7.—Neritic currents in the Rose Hill were able to create well-sorted siltstones with modal sizes as fine as .03 to .02 mm, which is essentially the lower size limit of quartz particles here. The close relation between modal size of the silt and clay content of the siltstones also illustrates sorting action.

8.—Good grain orientation and lamination in the Tuscarora estuarine and beach sands, and most of the Rose Hill neritic clay shale and siltstone beds shows a lack of burrowing organisms and probably a low faunal density. Swirled grain orientation, disturbed laminae, and irregular textural mottling shows an abundance of burrowers in parts of the Rose Hill and most of the Keefer lagoon-border sediments.

9.—The general paragenetic order among the cements is as follows: (1) calcite, ankerite, and authigenic tourmaline and illite, (2) hematite, (3) quartz overgrowths. Any of these may be lacking, and there is a little overlap in the paragenetic order; all these minerals have formed after deposition of the detritus and before jointing of the rocks. Furthermore, all these minerals have formed by direct precipitation into open pore space, not by replacement.

10.—Solution of detrital quartz grains under pressure to produce sutured and interlocking contacts is an important process in the Tuscarora and Keefer sandstones. Solution occurs in the Keefer only where clay films are present between the quartz grains.

11.—Iron is abundant in the Clinton strata (Rose Hill) because it was at this time that a plutonic igneous source became widely exposed in the source area. Iron was provided by intense weathering of the ferromagnesian silicate minerals in these primary rocks. Detrital magnetite survived in the oxidizing environments of the Rose Hill, whereas it was dissolved in the reducing environments. The bulk of the iron during Rose Hill deposition (and also in the red Tuscarora) was precipitated close to shore, presumably near the mouths of large rivers carrying it in from the heavily weathered source. In the shallow marine nearshore red sandstones the iron was oxidized and went into chlorite pellets and hematite cement;

here iron is most abundant. At moderate depths and distances from shore a mildly reducing environment prevailed, and the iron went into chlorite pellets and ankerite cement in the finer gray siltstones. Farther offshore very little iron was precipitated in the sediments and it all went into the highly reduced form of pyrite in the olive clay shales and greenish silt-shales.

12.—Patches of dark gray detrital clay have recrystallized in the Keefer sandstones to coarse water-clear flakes of illite and chlorite. The original detrital clay remains around the edges of these recrystallized patches.

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FIGS. 1–19, MARCH, 1960

ORIGIN OF POROSITY IN CARBONATE ROCKS[1]

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ABSTRACT

Porosity in carbonate rocks results from many processes, both depositional and postdepositional. An understanding of these processes and of the textural history of porosity is necessary to a full appreciation of the history of the rock and is important in the location of potential hydrocarbon reservoir rocks.

Several mechanisms appear particularly important in producing or changing porosity and pore-size distribution in carbonate rocks. Primary interparticle porosity is formed by deposition of a well-sorted calcareous sand or gravel under the influence of strong currents or waves or by local production of calcareous sand-size particles with sufficient rapidity to deposit particle on particle with little or no interstitial mud. Dissolution of interstitial mud in a calcareous sand can produce a microvuggy porosity resembling interparticle pore space. Simple cementation by calcite, anhydrite, dolomite, or quartz destroys porosity and pore size. Calcite cement appears to be especially common where the particles are monocrystalline, such as crinoid fragments. Primary constructional vugs are produced by formation of a rigid or semirigid framework. This framework may be organic or inorganic, and the interframework pockets may be filled with sediment or later with cement.

Sucrose dolomite contains the most common North American carbonate pore type. Available petrographic evidence indicates that it is formed by growth of randomly oriented, uniformly sized dolomite euhedra, accompanied by or followed by dissolution of the nonreplaced calcite. The interpretation of time and mechanism of dissolution depends on whether a local or a distant source of carbonate is assumed. Available evidence suggests that local source of carbonate and thus the development of porosity through dolomitization may be quite common. The intercrystalline pore geometry is produced by dissolution of the nonreplaced calcite. Thus only those limestones that have been converted to nearly pure dolomite by replacement and loss of nonreplaced calcite appear to be potential dolomite reservoir rocks. This porosity may be destroyed by calcite or by optically continuous dolomite cement. Limestone deposited originally as lime muds with abundant floating sand-size particles appears to be commonly preferred by the replacing dolomite euhedra.

The patchy growth of either scattered dolomite euhedra or masses of interlocking dolomite anhedra, followed or accompanied by dissolution of the nonreplaced limestone relict patches, produces vugs. A similar textural history of replacement by one mineral and dissolution of the nonreplaced patches is found in limestone replaced by fine networks of chert. Dissolution of the nonreplaced limestone produces porosity. Dolomitization of a limestone with interparticle or vuggy porosity without cementation of the earlier void space permits preservation of that void space.

Replacement by a mineral more susceptible to solution alteration, such as anhydrite, followed by dissolution of the replacement anhydrite appears to be a common vug-making mechanism. Vugs are also produced by simple dissolution of the more soluble particles or parts of particles, by fracturing, and by dissolution along fractures.

INTRODUCTION

The application of detailed petrographic study to understanding of the origin of pore space in carbonate rocks has received relatively little attention in the past. This has been true despite the economically important fact that a knowledge of the geologic history necessary for the formation of commercial carbonate reservoir rocks can be of considerable value in locating and predicting the areal distribution of such rocks. The need for such information becomes increasingly evident when it is realized that the location and geometry of many hydrocarbon accumulations in carbonate rocks are determined directly by the location and geometry of rocks with favorable capillary properties and perhaps only indirectly by structural factors. With such notable exceptions as Landes (1946), Imbt and Ellison (1946), Bergenback and Terriere (1953), Waldschmidt and others (1956), Graf and Lamar (1950), Towse (1957), and Edie (1958), very little has been written on the textural history of the negative part of carbonate rock fabrics, namely pore space.

During the present petrographic inves-

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tigations of North American carbonate reservoir rocks, 17 fields have been studied in detail. Twenty-two additional associated fields have been examined for supplementary and confirmatory data. These fields are distributed throughout the major carbonate-producing areas of western Canada, the Williston Basin, the Paradox Basin, west Texas, the midcontinent, and east Texas. Studies of both thin sections and polished surfaces were utilized in conjunction with mercury capillary pressure measurements. Such measurements provide information on pore-size distribution in the rock and are useful tools for observing changes in pore fabric with changes in rock fabric or composition. In addition, mercury capillary pressure data provide a useful measure of the value of the rock as a potential reservoir. This paper discusses some of the observed mechanisms and processes that appear to be significant in changing pore volume and pore-size distribution in carbonate reservoir rocks.

METHODS

Permeability plugs drilled at approximately 1- or 2-ft intervals along the sides of cores have provided the basic source of study material. These cores were obtained from hydrocarbon-productive rocks, as well as from nonproductive associated rocks. Porosity, air permeability, and capillary pressure data of each plug were determined, and a thin section was cut from one end. Because the long axis of the plug as cut is perpendicular to the long axis of the core, the thin section is approximately perpendicular to the bedding if the strata are horizontal. Such thin sections permit direct comparison of the petrophysical (Archie, 1952) properties obtained from the rock with the petrology. These observations were commonly supplemented with examination of polished surfaces with a binocular microscope and with a metallographic microscope. Additional data were obtained from chemical analysis, X-ray, and replica peels.

Certain techniques have been especially useful in studying carbonate thin sections; for example, thin sections prepared without cover glass can be lightly etched on one corner with diluted acid which brings the dolomite, anhydrite, and quartz into relief and facilitates examination or quantitative estimation. With more extensive etching, all the carbonate in one spot can be removed, leaving the residue exposed for examination (see Lees, 1958). If the uncovered section is to be photographed or examined under high magnification, the section can be covered with a glycerine-water mixture and temporarily placed under a cover glass. Low-power (×6–40), wide-field, stereopolarizing microscopes have proved very useful for observing most carbonate rock and pore fabrics. Standard petrographic microscopes were used for detailed examination of individual grains and crystals. Such examination is particularly critical for understanding how grains or crystals fit together and thus for determining interparticle or intercrystalline pore relationships.

The mercury capillary pressure curve (Purcell, 1948) is a useful device for understanding and illustrating pore-size distribution. Mercury, a nonwetting fluid, is forced into the rock sample under increasing pressure. Because the pressure required to inject mercury into a pore is a function of the entry radius of that pore, the curve of cumulative volume injected plotted against injection pressure is related to the distribution of entry sizes in the rock. The system of largest interconnected pores will be entered at the lowest pressures, and with increasing pressure smaller pores will be entered. For example, a pore entry 0.002 mm in diameter will be entered by mercury at approximately 107 psi. The mercury-injection curve thus provides an indication of the ability of the rock to hold and transmit fluid. This is true because permeability is related to the size, amount, and degree of interconnection of the larger pores. Because oil is normally the nonwetting fluid in an oil-water system, there is a relationship between the ratio of oil to water to be expected in a rock at a given height above the free-water level and the amount of mercury injected at a pressure equivalent to that height. Thus, capillary pressure data provide direct information on the potential value of a rock as a reservoir rock under any given conditions. From the capillary pressure data it is possible to isolate these pore types that contribute significantly to hydrocarbon production from associated porous rocks with fine-pore interconnection.

It is interesting to note that, particularly

in carbonate rocks, many of the larger pores or vugs are entered at pressure far in excess of the pressure that would be expected for their size. This is true because mercury entering such pores must first pass through pore channels of much smaller size. Despite this limitation, the capillary pressure curve provides specific information on the capillary properties of the rock and is of considerable use to the petrographer interested in porosity development. Its use is somewhat analogous to the use of grain-size-distribution curves in studying noncarbonate clastic sediments.

POROSITY DEVELOPMENT

Experience gained thus far in these studies has shown that most observed pore types can be grouped for convenience into three general categories: (1) primary, (2) sucrose dolomite, and (3) secondary vugs. Primary porosity includes those pore types which result from depositional processes and which have later been modified only slightly by compaction, pressure solution, simple pore-filling cement, or dissolution alteration. In general, the original pore fabric is intact and recognizable. Sucrose dolomite has a mutually interfering texture composed of euhedral to subhedral crystals which are only slightly welded at points of contact. Such a texture necessarily results in intercrystalline pore space and a rock surface resembling that of lump sugar. Secondary vugs are void spaces formed by postdepositional processes; they are larger than would normally be expected from the simple fitting together of associated mutually interfering crystals or deposited particles.

These general categories by no means form a suggested classification but only provide a framework for examining processes and mechanisms of porosity formation and destruction.

Primary Porosity

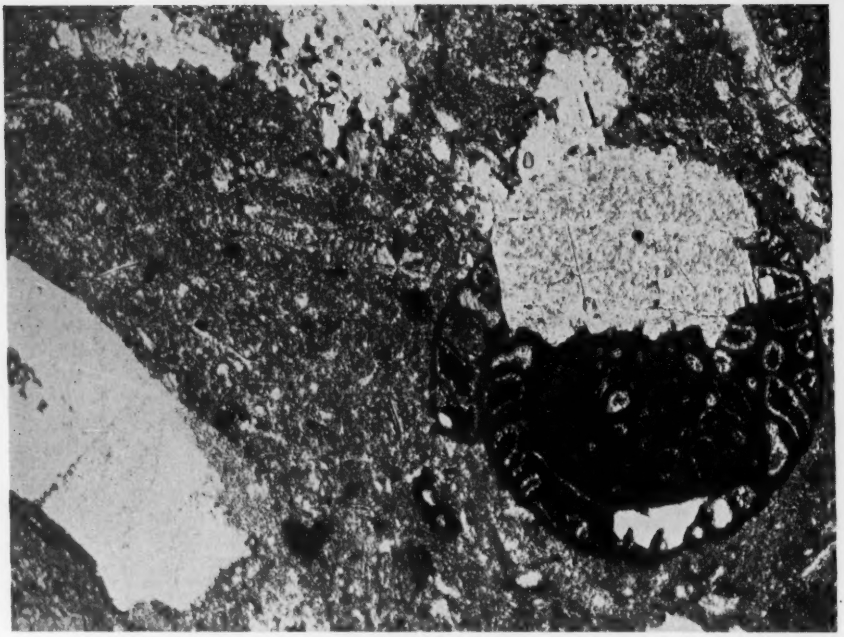
Carbonate primary pore types originate from the three common carbonate sediment types—framework, muds, and sands.

Framework.—Under framework are included those carbonates which, by the nature of the deposit, were originally bound into a rigid or semirigid framework. This framework material may be organic or inorganic and is commonly composed of tightly interlocking crystals of calcite or aragonite. Most significant is the fact that such a famework is often capable of allowing the formation of primary constructional vugs within the frame. Vugs are formed when the frame is incompletely filled with finer sediments. These frameworks are seen in some coral reefs, in oyster banks, and in rudistid banks; they can be produced by prolific growth of some sponges, calcareous algae, hydrocorallines, and, perhaps, bryozoans. In general, framework, as used here, would include the rigid framework of "true reefs" and some nearly "true reefs," as discussed by Lowenstam (1950). On a microscale, the relatively unimportant pore contribution from open chambers formed by removal of organic matter from within some fossils should also be included (fig. 1). In addition, most hot spring and some cave deposits represent framework. In these deposits a framework of rigid, interlocking crystals deposited either organically or inorganically commonly grows in such a way as to leave empty spaces or voids.

Examples of constructional void porosity in North American carbonate reservoirs are extremely rare.

Mud.—One of the most commonly observed carbonate rock types was deposited as fine lime mud (fig. 2). The origin of the contributing particles is usually not determinable, and the particles may be chemically or biochemically precipitated fine crystals or finely comminuted debris from originally larger particles. The distinguishing characteristic of this group is its particle size, commonly less than 10 μ, and thus the small size of its interparticle pores. Like its terrigenous clastic counterpart, shale, this rock under certain conditions may retain relatively high porosity, but the fine pore size commonly excludes it as a commercial reservoir rock. Unlike shale, the carbonate mud appears especially susceptible to pressure solution and simple interparticle cementation, resulting in reduction in porosity and pore size (fig. 1).

Sand.—Carbonate sediments deposited as sorted sands or gravels form the third group (fig. 3). Here again, the analogy can be drawn to the terrigenous clastic sediments. These sediments were deposited, in general, under conditions of sufficiently rapid water movement (that is, waves, tidal currents, or

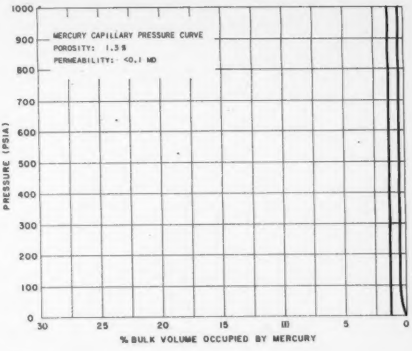


Crossed nicols, ×30.

FIG. 1.—Carbonate mud with floating fossil fragments—cemented. This rock was deposited as a carbonate mud with floating fossil fragments. The mud has since been compacted and cemented with accompanying reduction in porosity and pore size. Vugs representing original chambers within the fusulinid are present but are interconnected only through the fine-pore matrix.

Replacement anhydrite (white) has partially replaced the fusulinid and has replaced the crinoid fragment. The tendency of the anhydrite to conform to the particle being replaced and the tendency to maintain the rectangular outline both within and outside the particle are apparent.

Desert Creek field, Utah; Paradox Formation (Pennsylvanian).



longshore currents) either to remove the finer particles or to transport and selectively deposit the larger particles. One important difference between terrigenous clastic particles and carbonate particles is the fact that many of the carbonate particles are produced locally, that is, at or relatively near the site of deposition, and their size distribution may often reflect the type of particle available as well as the mechanical sorting. For example, it is not uncommon to find a relatively large shell or cluster of shells within fine mud. Alternatively, excellent sorting may be achieved by produc-

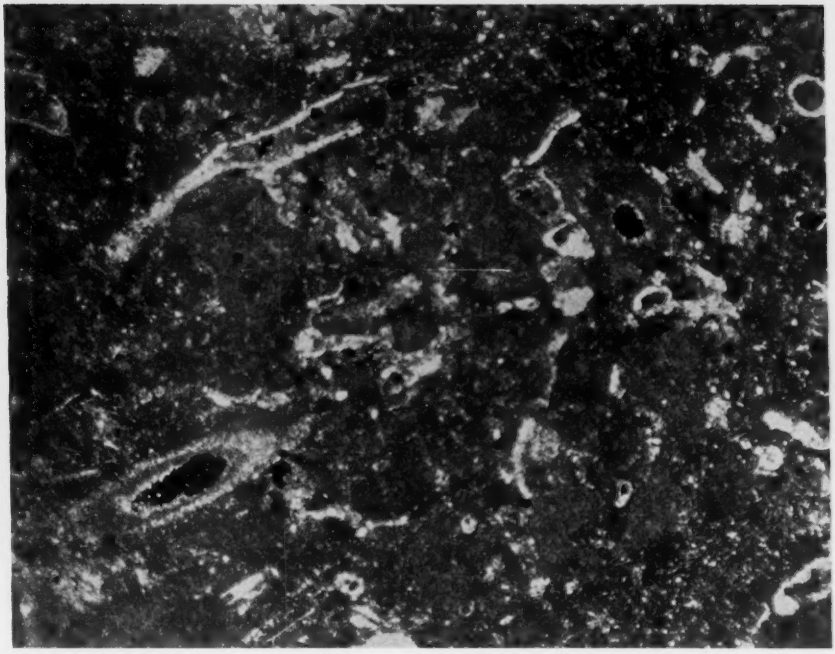


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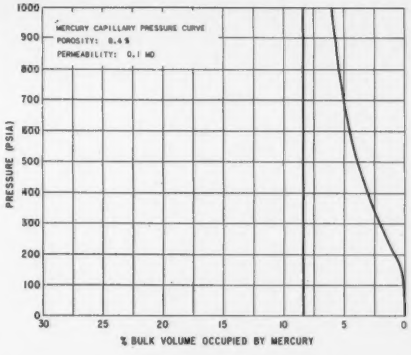
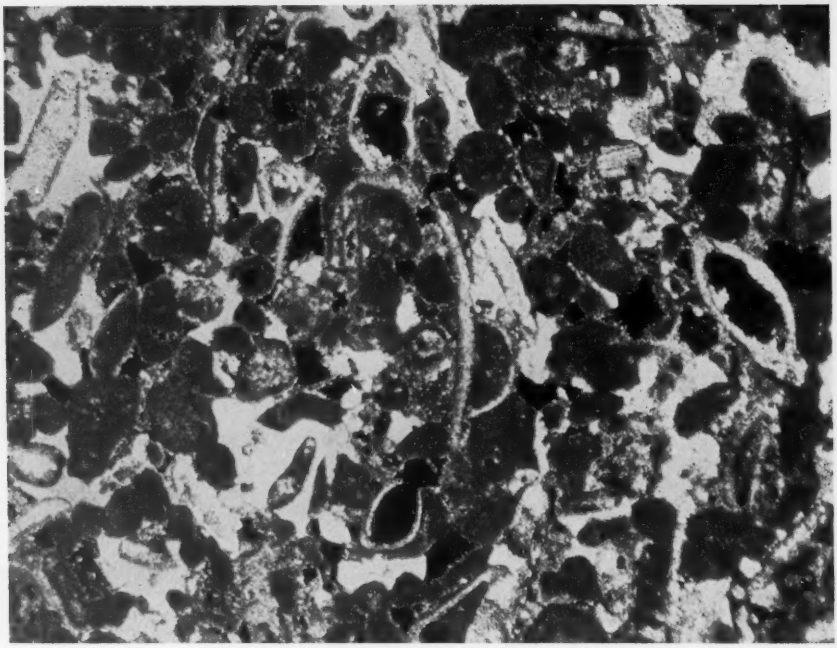
FIG. 2.—Carbonate mud with abundant floating sand-size particles. This rock was deposited as a carbonate mud with abundant floating sand-size pellets and fossil fragments. The fine, light specks are 5–7 μ dolomite euhedra (20 percent of the rock) replacing the original carbonate mud. The large vug in the upper right has been interpreted as the interior of a productid spine and is connected only through the very fine pores of the partially cemented and compacted mud.

Middle field, Saskatchewan; Charles Formation (Mississippian).



tion of only one type of particle such as pellets or crinoids. The nature of such particles is extremely varied, largely because of the important biologic contribution. Disruption and mixing of sediments of all sizes by burrowing organisms is also common. The nonskeletal particles, such as oolites, pellets, bahamite grains (Beales, 1958), and algal accretionary structures, present special problems. Determination of the origin or even the recognition of genetically similar or dissimilar objects in this group is difficult and often impossible with present criteria.

Commonly included among sand-size and larger particles are whole or broken fossil animal and calcareous plant fragments,



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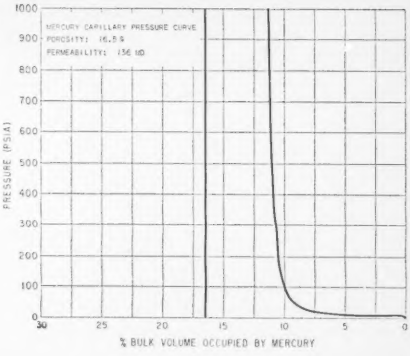


FIG. 3.—Primary interparticle porosity in a sorted, calcareous sand. This rock is composed of well-sorted pellets and some fossil fragments. The porosity is interparticle, and calcite cement has filled some of the pore space. This cement has grown as simple sparry anhedral crystals and as optically continuous growth on crinoid fragments (upper left). The pore space that was not filled by mercury at 1000 psi is probably within the pellets whose clay-sized carbonate particles have been incompletely cemented.

Midale field, Saskatchewan; Charles Formation (Mississippian).

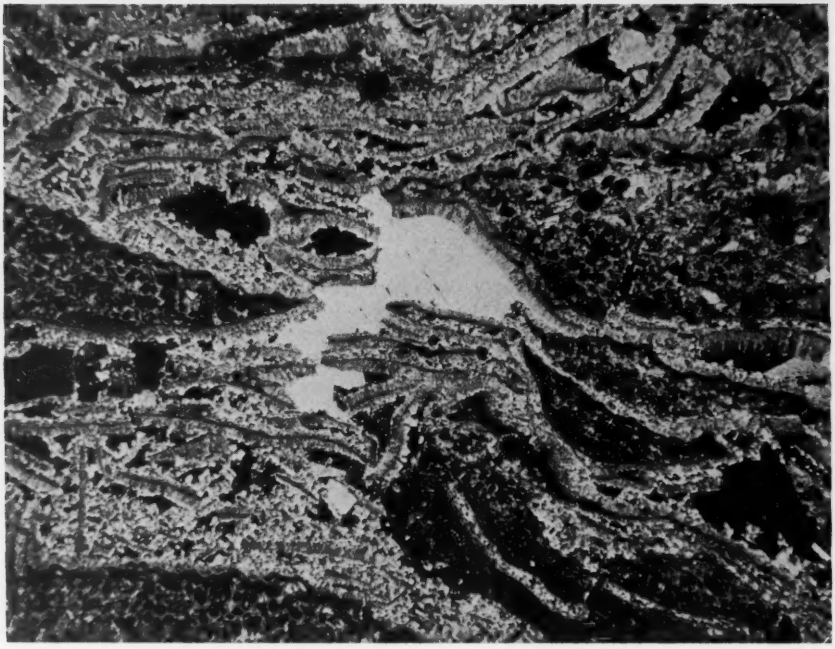
oolites, pellets (both fecal and bahamite grains), grapestone lumps (Illing, 1954), algal accretionary bodies, lumps of partially lithified contemporaneous sediment, and some fragments of older carbonate rocks. When free of mud these carbonate sands have porosities and pore-size distributions similar to their well-sorted, terrigenous clastic cousins. However, the carbonate sands are often poorly sorted. In fact, a complete physical gradation exists from sorted carbonate sands to those with a carbonate mud

matrix between self-supporting sand-size particles to those with abundant sand-size particles floating within mud to mud with few, if any, sand-size particles. Naturally, the addition of the mud matrix reduces the porosity and pore size in primary sediments.

Because of local production of sand-size and larger carbonate particles, sediments with varying amounts of carbonate mud matrix or varying amounts of sand-size particles within mud are possible. Indeed, rocks deposited in a mud environment often show a patchy distribution of apparently well-sorted sand-size particles or sand-size particles with the interstitial spaces incompletely filled with mud. These can result from what we might term competitive sedimentation; that is, the final sediment is a product of two competing rates—rate of particle production and rate of mud deposition. The rate of mud deposition is of course determined by the rate of production of silt and clay size carbonate particles or crystals and the ability of currents to selectively remove mud from the site of deposition. The production and deposition of sand-size particles are sometimes sufficiently rapid to keep pace with mud sedimentation. At times production and deposition of particles is so rapid that the larger carbonate particles fall one on top of the other, producing interparticle void space. This space may or may not contain some mud. Plate-shaped calcareous algae (Parks, 1958) are ideally suited for this type of competitive sedimentation (fig. 4). However, crinoids, fusulinids, and other particles are sometimes deposited either originally or through borrowing with partially mud-filled interparticle pores showing crude geopetal structures (Sander, 1936). The essentially in place accumulation of sand-size particles with little or no interstitial mud appears well illustrated by the deeper lagoon sediments of Kapingamarangi Atoll (McKee, and others, 1959).

Within the framework group, the bulk of the sediment exclusive of the frame is commonly composed of carbonate sands and mud. Indeed, primary-constructed vugs within the frame are often observed to be filled or partially filled with fine particulate carbonate mud. Where this vug sedimentation has been incomplete, a geopetal structure or floored cavity is produced and some primary vug porosity is retained.

Cementation.—Each of these primary pore types may be destroyed; porosity and pore size may be reduced either by cementation or by cementation in conjunction with replacement. Commonly observed cementing materials are calcite, anhydrite, and quartz. Some carbonate cements, such as those found in beachrock (Ginsburg, 1953), may be formed very early, and some may be formed much later. The age of most cements is not determinable with present criteria. However, the relative age of two or more cements may sometimes be determined petrographically. These cements may simply grow into previously open voids (fig. 5), or by a replacement process they may engulf areas of fine carbonate mud with its contained fine interparticle porosity. In either case, the cement may or may not be an optically continuous overgrowth on a monocrystalline carbonate fragment, such as a crinoid columnal (fig. 6). By either process, coarse sparry calcite may result. An excellent critical statement of these processes and the criteria for their differentiation have recently been published by Bathurst (1958). Bathurst recognizes two types of cements, depending on whether the cementing material grew into void space or replaced carbonate mud. The cements growing into interparticle voids may be optically continuous on single crystal particles such as crinoid fragments (rim cement); they may produce small crystals commonly oriented perpendicular to the void walls (granular cement). Some of these crystals may grow preferentially to form a coarse mosaic. Cements occupying spaces previously occupied by carbonate mud (grain growth) may grow by replacement in optical continuity with a large host, such as a crinoid fragment, or by grain enlargement of several of the small particles to produce a coarse mosaic. Bathurst presents criteria for recognizing and interpreting these diagenetic fabrics and makes a much-needed plea for the elimination of the term "recrystallization" in favor of specific recognition of the individual processes. Reduction in porosity by pressure solution appears to be an important mechanism in carbonate rocks. This process reduces porosity and pore size and produces



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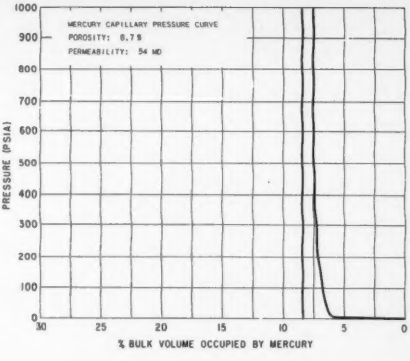


FIG. 4.—Primary porosity—competitive sedimentation. This pore type was formed by cosedimentation of pelleted carbonate mud and plate-shaped calcareous algae (*Ivanovia*, Khvorova, 1946). Production of the alga was sufficiently rapid to permit leaf to fall on leaf with little deposition of interstitial mud. Thus, pore space (black) is found between the algal plates. Cementation by fibrous calcite lining the interplate pores, by pore-filling coarse calcite (center), and by anhydrite, along with cementation and compaction of the pelleted mud, has reduced the porosity. Crumbling in place of the fragile plates has caused some local brecciation in both plates and mud. The pelleted mud lens deposited on an algal plate (left center) has a slightly rippled uppersurface.

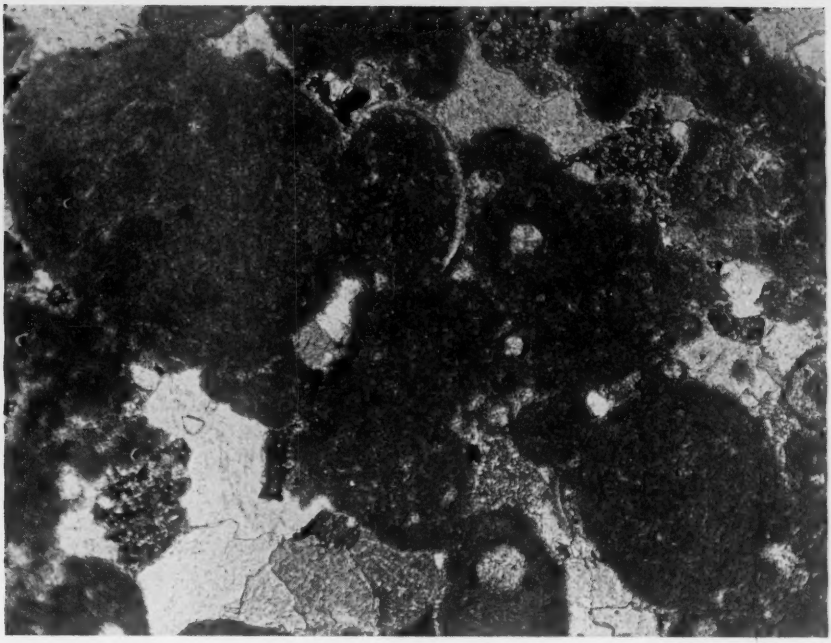
Most of the mercury is injected at very low pressure, indicating the interconnection of the void space. Most of this void space is visible.

Desert Creek field, Utah; Paradox Formation (Pennsylvanian).

an interlocking texture. Indeed most particulate carbonate rocks with little mud exhibit a particle fit that is much tighter than could possibly be produced by simple sedimentation.

Sucrose Dolomite

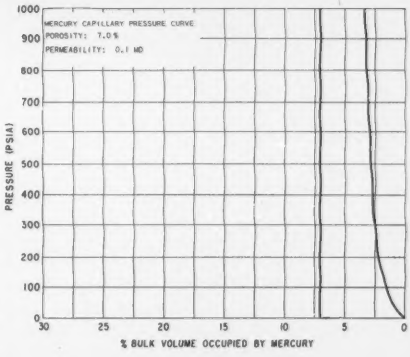
Quantitatively, the most important North American carbonate pore type in terms of oil and gas production, and one of the most genetically interesting, is found in sucrose



Crossed nicols, ×70.

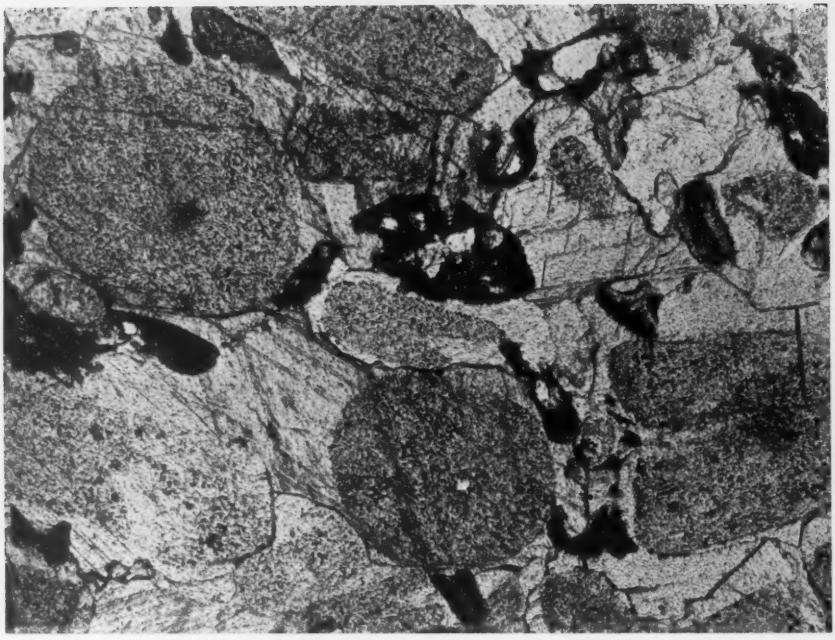
FIG. 5.—Primary interparticle porosity in a sorted, calcareous sand—cemented. This rock, similar to fig. 3, is composed of well-sorted pellets and some fossil fragments. The original interparticle porosity has been almost completely cemented by simple sparry calcite anhedra. The pore volume that was not filled by mercury at 1000 psi is probably within the pellets of fine carbonate mud.

Midale field, Saskatchewan; Charles Formation (Mississippian).



dolomite (figs. 7 and 8). The crystal size ranges from less than 5 μ to more than 100 μ with 25–50 μ being very common in reservoir rocks. The specific relation of porosity to dolomitization has long been debated (Hohlt, 1948, p. 26), and the problem was recently summarized by Fairbridge (1957, p. 159).

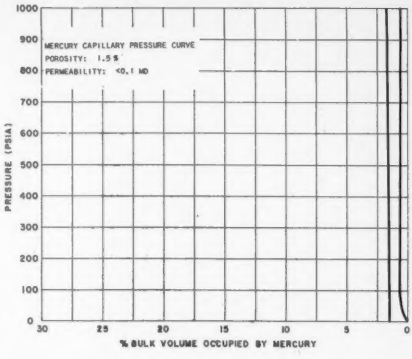
Petrographic evidence for the origin of porosity in sucrose dolomite comes from areas where all degrees of dolomitization are present in intimate vertical and horizontal association. One such example is the Mississippian Charles Formation, "Midale beds," in the Midale field, Saskatchewan (Parker, 1956, and Edie, 1958). In a section approximately 50 ft thick carbonate rocks exhibiting all degrees of dolomitization are observed. The highly dolomitized limestone generally is concentrated in the upper part of the section (fig. 7). In addition, the orig-



Plane polarized, ×33.

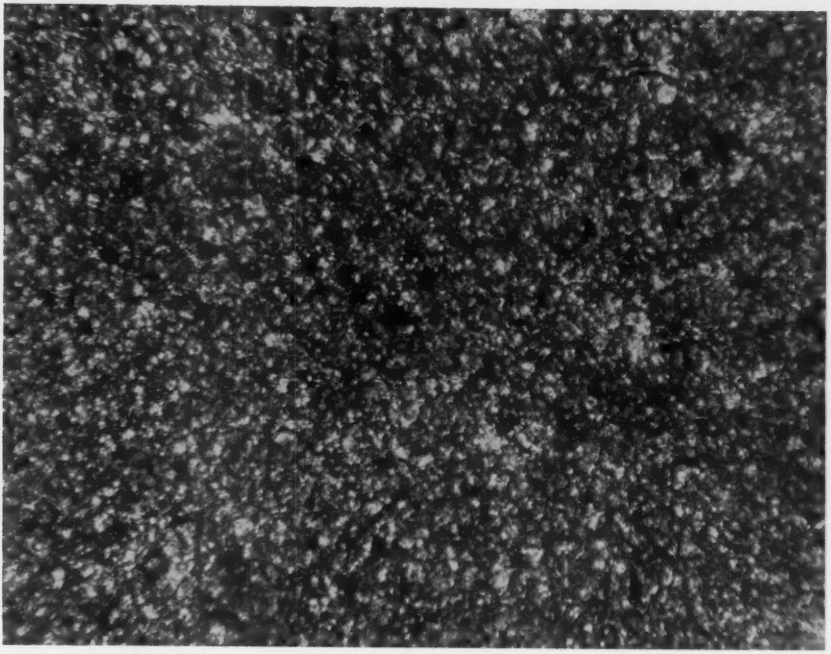
FIG. 6.—Well-sorted crinoidal and bryozoan carbonate sand—cemented. This carbonate sand has been completely cemented by optically continuous overgrowth on the crinoid fragments. Such cementation is common where the particles are single crystals.

Andrews South Devonian field, west Texas; Subsurface Devonian.



inal limestone shows all gradations from sorted carbonate sands to muds with varying amounts of contained carbonate sand-size particles (figs. 2 and 3). Observed sand-size particles include crinoid parts, ostracods, algal balls, and pellets. The following petrographic and petrophysical evidence from these Midale field rocks is applicable to the origin of this porous texture:

1.—Carbonate rocks with dolomite contents of from 3 to 90 percent of total solids are present. In the slightly dolomitized limestone the dolomite occurs as scattered perfect rhombs (fig. 2). These rhombs are of replacement origin and appear to occupy volumes once occupied by parts of pellets, fossil fragments, fine carbonate mud, and cross boundaries between these elements. As the dolomite content increases, the number of rhombs increases. In general, the rhombs in an individual rock sample are all approxi-



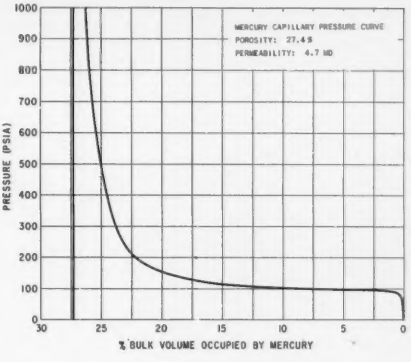
Crossed nicols, ×200.

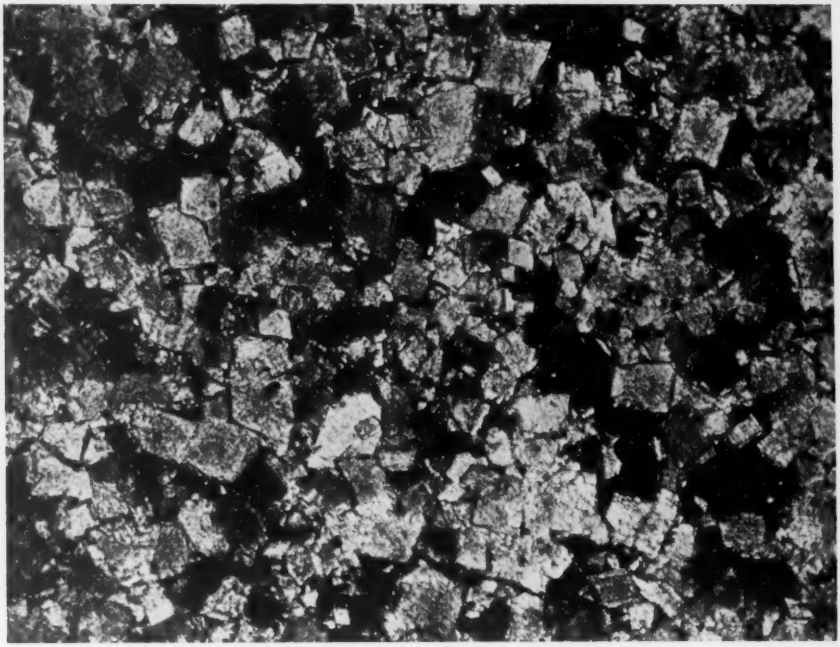
FIG. 7.—Fine sucrose dolomite. This uniform, fine sucrose dolomite is composed of randomly oriented, mutually interfering dolomite euhedra of approximately the same size, 5–7 μ. The porosity is wholly intercrystalline and analogous in pore-size distribution to a well-sorted siltstone. This porosity was formed by replacement of limestone by the dolomite rhombs, followed or accompanied by dissolution of the nonreplaced limestone. The nature of the original limestone is not determinable from this thin section. However, study of associated partially dolomitized limestones indicate that the rock being preferentially dolomitized began as a lime mud with abundant sand-size particles (see fig. 9).

Versenate chemical analysis:

Dolomite	86 percent
Calcite	8 percent
Anhydrite	0 percent
Residue	6 percent

Midale field, Saskatchewan; Charles Formation (Mississippian).

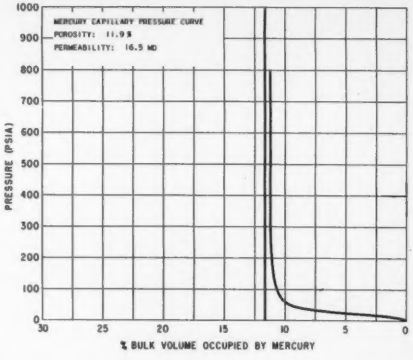




Crossed nicols, ×77.

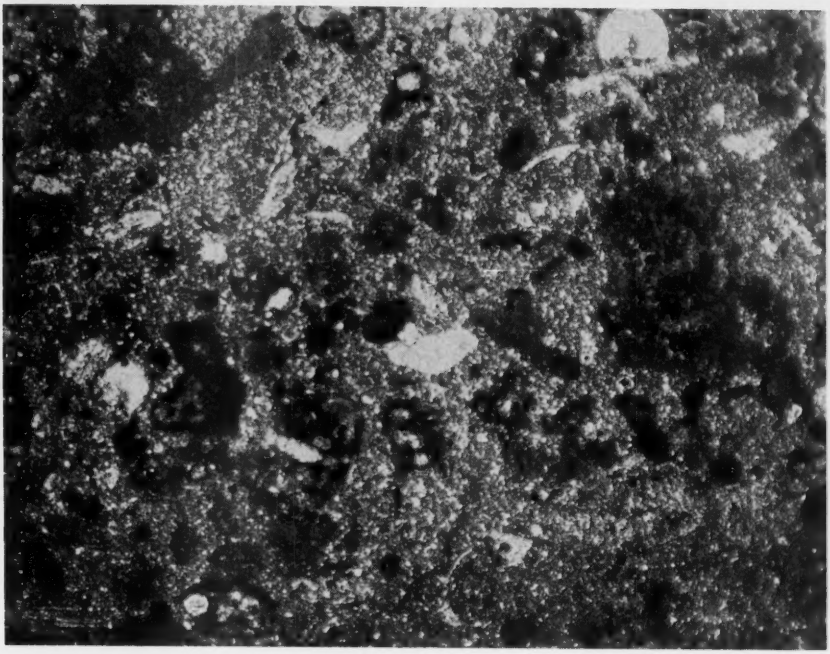
FIG. 8.—Sucrose dolomite. Euhedral dolomite crystals, 50–75 μ in diameter, have developed an excellent mutually interfering texture with intercrystalline porosity. The porosity was formed by growth of the rhombs, followed or accompanied by dissolution of the nonreplaced limestone. There appears to have been some minor addition of dolomite in optical continuity, which slightly reduced porosity and pore size. (Compare with fig. 13.)

Gas City field, Montana; Red River Formation (Ordovician).



mately the same size, 5–20 μ, and appear to show little, if any, preferred orientation. At approximately 50 percent dolomite the relatively large rhombs have obscured most of the original very fine calcite mud and particles, leaving only small patches of the original texture and a cloudy background of fine calcite with faint ghosts of particles (fig. 9).

2.—The plot of porosity versus dolomite content for these same rocks indicates that there is a slight decrease in porosity up to approximately 50 percent dolomite (fig. 10). As the dolomite content increases above 50 percent, the porosity rapidly increases to values approximating 30 percent at 80 to 90 percent dolomite. It is commonly observed in dolomitization sequences that early



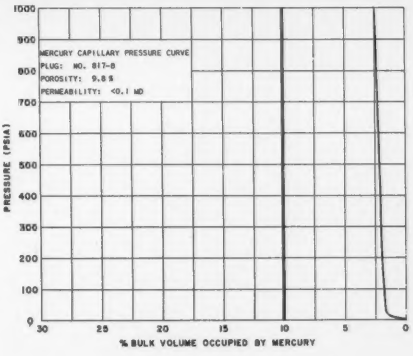
Plane polarized, ×35.

FIG. 9.—Carbonate mud with abundant floating sand-size particles—37 percent replacement dolomite. This rock was deposited as a carbonate mud with abundant sand-size skeletal and nonskeletal particles. Growth of small (approximately 7–10 μ) dolomite euhedra within both mud and particles has obliterated much of the original texture. Note the low porosity and small pore size.

Versenate chemical analysis:

Dolomite	37 percent
Calcite	56 percent
Anhydrite	0 percent
Residue	10 percent

Midale field, Saskatchewan; Charles Formation (Mississippian).



stages of dolomite growth are accompanied by decreased porosity and reduced pore size. Porosity and pore size increase as complete dolomitization is approached.

These data must be interpreted in the light of two possible assumptions: (1) Carbonate as well as magnesium is brought to the site of rhomb growth in solution from other rocks (distant source). (2) Carbonate is derived locally from the rock being dolomitized during dolomitization (local source).

If the new dolomite crystals are grown from solutions that are bringing carbonate

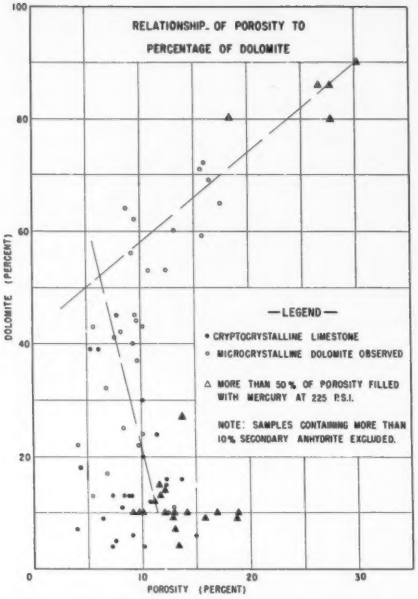

RELATIONSHIP OF POROSITY TO
PERCENTAGE OF DOLOMITE
DOLOMITE (PERCENT)
POROSITY (PERCENT)
—LEGEND—
• CRYPTOCRYSTALLINE LIMESTONE
○ MICROCRYSTALLINE DOLOMITE OBSERVED
△ MORE THAN 50% OF POROSITY FILLED WITH MERCURY AT 225 P.S.I.
NOTE: SAMPLES CONTAINING MORE THAN 10% SECONDARY ANHYDRITE EXCLUDED.

FIG. 10.—Relationship of porosity to percentage of dolomite. These samples were taken from the "Midale beds," Charles Formation, Midale field, Saskatchewan. The dolomite was obtained by Versenate analysis, and the porosity was determined by the standard grain-density method. Points outlined with a triangle are expected to be capable of commercial hydrocarbon production. The triangle-outlined points in limestone are generally sorted pelletoid-bioclastic sands with interparticle pore space (fig. 3). These should be excluded from consideration when the data are used for an understanding of the formation of sucrose dolomite from a lime mud with abundant sand-size particles. The high-porosity reservoir rocks are uniform, 5–7 μ, sucrose dolomite (fig. 7).

into the rock being dolomitized, the porosity should decrease. Most of the solid dolomite rhombs are in effect occupying space once occupied by a multitude of much finer cryptocrystalline calcite grains with their very fine intergranular porosity. If the porosity of the volume being replaced by one dolomite crystal is less than approximately 43 percent, there will be sufficient calcium to combine with the added magnesium to produce the dolomite. Carbonate must be added to maintain the volume-for-volume growth. The increase in porosity above 50 percent dolomite, with increasing dolomite content but similar dolomite rhomb size and texture, indicates that the original calcite matrix between the dolomite rhombs has been dissolved to produce void space. Dissolution of this relict calcite following dolomitization is necessary if the carbonate has been added to the rock from an outside source. These changes in pore geometry are further substantiated by changes in pore-size distribution, as indicated by capillary pressure measurements. With increasing dolomite, the pore size increases. Undoubtedly, some small amount of dolomite could have been added both as new rhombs and as enlargement of the earlier rhombs during the later stages of evolution of the present Midale dolomite texture. The result is high porosity, uniform pore size, and 4–10 md permeability in the relatively pure dolomite (fig. 7). The uniform pore size is to be expected from the uniform rhomb size. Similar textural changes are recognized by Barnetche and Illing (1956) in the Tamabra Dolomites. The porosity and permeability cannot be produced by addition of more dolomite but only by loss of interstitial calcite, which unavoidably results in concentration of dolomite. Murray (1930) recognized dissolution of calcite from a dolomitic limestone as the origin of dolomite porosity, but he considered selective leaching at an unconformity by meteoric waters as the mechanism. Dissolution of the relict calcite in the Midale rocks could have taken place during at least two possible times in the history of the rock: (1) during the late stages of the dolomitizing process by excess solution over dolomite growth, as proposed by Landes (1946), or (2) during the pre-Jurassic erosion period by meteoric water. The latter appears unlikely because of little evidence for widespread dissolution in the adjacent highly porous and permeable limestone units. Dissolution appears restricted to those dolomites which had already attained a high degree (> 50 percent) of scattered rhomb dolomitization.

If the assumption is made that all carbonate used in the growth of the dolomite is derived locally from the rock being dolomitized and is relatively near the site of rhomb growth, an alternative explanation of the Midale and similar data is possible (Weyl, 1960). Assume that the carbonate mud with abundant floating sand-size carbonate par-

ticles had approximately 21 percent porosity at the time of dolomitization. The growth of a dolomite rhomb would involve (1) addition of magnesium from outside the system and (2) dissolution of some calcite from outside the volume of rhomb growth to provide the necessary carbonate for filling the pore space within the volume of rhomb growth plus the 12–13 percent porosity difference occasioned by growth of the denser mineral dolomite. Excess calcium ions must be removed from the system. If this process is carried to completion, all the calcite will be used to form dolomite. The resulting rock would have a porosity, assuming no compaction during and following dolomitization, equaling the sum of the original porosity plus an increase of 12–13 percent of the calcite originally present. In the Midale example an original porosity within the calcite mud and particles of approximately 21 percent would result, by the above process, in a porosity of approximately 30 percent, assuming no compaction during or after dolomitization. This pore space would be between the rhombs; that is, the rock would have a sucrose dolomite texture. If compaction took place during or after dolomitization, a higher original porosity would be necessary to produce the observed 30 percent porosity. However, in all stages of partial dolomitization the rock would appear as limestone with contained dolomite euhedra.

If insufficient carbonate (CO_2, HCO_3^-, CO_3^{--}) is available, calcite dissolution must accompany growth of dolomite, thus forming and redistributing porosity (local source). If sufficient carbonate is brought to the site of dolomitization (distant source) then porosity destruction will occur. In the latter case, only later calcite dissolution, possibly occasioned by subaerial exposure, can produce dolomite porosity. This may be an important distinction in terms of dolomite porosity location and distribution. If dolomite porosity can be formed only through simple calcite dissolution after partial dolomitization, then those partially dolomitized limestones that have been subaerially exposed should have the best chance for porosity development. However, if porosity development is part of the dolomitizing process, then all pure and nearly pure dolomites should be porous regardless of their history or nearness to unconformities if that porosity has not been destroyed by compaction pressure solution and/or simple cementation. The amount of porosity should depend on the porosity of the limestone at the time of dolomitization, assuming no excess calcite dissolution and no compaction, pressure solution, or cement.

One observation that has a bearing on this problem is the recognition that in the author's experience dolomite commonly grows by replacement as opposed to growing as cement. In a sediment composed of carbonate mud and sand-size particles the mud is commonly replaced first. This may or may not be followed by replacement of the sand size particles in their order of susceptibility to replacement. In mud-free carbonate sand, either skeletal or nonskeletal, dolomite is commonly observed to replace the particles rather than growing as cement in the interparticle areas. In such cases there is the ever present possibility that the interparticle voids were cemented with sparry calcite which resisted dolomitization and was later dissolved. Where dolomite is observed between apparently mud-free carbonate sand particles, it may have replaced small amounts of interparticle carbonate mud or earlier calcite cement. However, oolites are commonly dolomitized with apparent retention of the interparticle void space.

Dolomite cement in limestone as veins and vug filling appears to be exceptionally rare. Indeed porous quartz sandstones interbedded with dolomite such as the St. Peter Sandstone in southwestern Wisconsin, appear to contain little true dolomite cement despite the fact that carbonate mud, sand, and gravel above (Plattville) and below (Prairie du Chien Group) have been completely dolomitized. Presumably the waters which caused the growth of dolomite by replacement in at least the overlying limestone were common, in general, to both the limestone and adjacent sandstone. However, minor compositional differences could have been caused by differences in the nature of the rocks. Where dolomite cement is observed in quartz sandstones, the dolomite may have grown at the expense of earlier calcite.

Calcite and anhydrite cements are extremely common in both limestone and dolomite. Although both are commonly ob-

served as replacement crystals, free growth as cement appears to be preferred. Indeed, the replacement crystals often begin by free growth in voids. Simple calcite cement implies gross transport of carbonate.

If the above generalizations are true (that is, dolomite in normal sedimentary rocks commonly grows by replacement rather than cement as opposed to the other two common rock carbonate minerals, calcite and anhydrite, which appear to grow as cement), they offer some additional insight into dolomite porosity development. Perhaps the preference for replacement indicates that the growth of dolomite commonly required carbonate (local source) and calcite had to be dissolved in order to provide carbonate for dolomite to grow. The growth of the dolomite takes place simultaneously with, and at the site of, maximum calcite dissolution, thus is replacement. Thus waters from which dolomite commonly grows may be similar to many present day subsurface waters in that the ratio of Mg and Ca to total CO_2 is large.

Exceptions to the above generalizations, such as dolomite veins in limestone or other forms of dolomite cement, occur in association with metallic mineral deposits. In such cases there may actually have been gross transport of carbonate to the site of dolomite growth. The validity of the generalizations, however, should be tested by additional observations of replacement and free growing dolomite.

If it is assumed that on further compaction the partially dolomitized calcitic mud compacts whereas the sucrose dolomite does not compact and therefore retains its intercrystalline porosity, the following explanation of figure 10 is possible: The limestones with less than approximately 50 percent dolomite compacted and were cemented to their present porosity. At approximately 50 percent dolomite or slightly greater, depending on the actual uniformity of the replacing rhombs, the dolomite crystals would effectively act as a frame and would inhibit further compaction. Rocks with more than 50 percent dolomite would be expected to retain some porosity, and the amount would increase with increasing dolomite content. The pore size should also increase with increasing dolomite, because there would not be enough relict calcite to fill the interrhomb space.

The concept of local derivation of the necessary carbonate ions appears to offer a more logical explanation for the Midale data. However, determining the time of and the mechanism for removal of calcite that previously existed in the intercrystalline spaces of sucrose dolomite remains one of the most critical problems in understanding the origin of dolomite porosity. Thus, dolomite porosity results from dissolution of nonreplaced calcite either following or as part of the dolomitizing process, depending on whether carbonate is added to the rock being dolomitized or is derived locally by dissolution of calcite.

It should be emphasized here that in invoking this origin of the sucrose texture, it has been assumed that, because all stages in the process can be seen in associated rocks and because the porosity and permeability data conform to the observed change, there is a genetic sequence from fine particulate limestone to sucrose dolomite. The mutual similarity of dolomite rhombs, whether they occur as scattered crystals in the limestone matrix or as crystals in the sucrose texture, argues for the above assumption. In addition, the cryptocrystalline calcite matrix observed in the relatively pure limestone appears to be identical to that found as relict patches in incompletely dolomitized rock. However, the relatively pure, high-porosity dolomite gives little evidence in itself of having been derived from a cryptocrystalline limestone. Evidence that these high-porosity dolomites are end members in a genetic sequence is supplied only by associated rocks of lower dolomite content and by the tendency of rocks with more than 50 percent dolomite to increase in porosity with increase in dolomite content. This increase occurs with observed loss of calcite and with no apparent change in dolomite texture.

It is interesting to note that in the Midale field example the preferentially dolomitized limestone was the lime mud with contained floating sand-size particles rather than the associated higher-permeability sorted carbonate sands. In addition, the insoluble residue in the limestones, many of which contain little or no carbonate mud matrix, averages less than 2 percent and seldom ex-

ceeds 3 percent, whereas the residue in those rocks with greater than 20 percent dolomite varies from 2 percent to 18 percent (fig. 11). This residue is largely silt- and clay-sized quartz, which would be expected to be concentrated in the fine particulate sediment. The possibility that the increase in residue results from concentration during dolomitization by thinning of the section is excluded, because a reduction to approximately 1/10 of the original limestone thickness would be required. This higher residue of fine derived particles further suggests the preferential dolomitization of the quiet-water carbonate mud by scattered dolomite euhedra.

Additional evidence for the origin of porosity in dolomite is found in examples of sorted crinoidal sands and laterally equivalent crinoidal sands with carbonate mud matrix. One example of such an association is found in the Mississippian Turner Valley Formation in the Harmattan area, Alberta. Here, as is quite common with such sediments, the originally well-sorted crinoidal sand has been cemented by optically continuous calcite overgrowths, and it remains limestone. The poorly sorted associated carbonate sands have been dolomitized with scattered rhombs, followed or accompanied by dissolution of the nonreplaced calcite to produce pore scace (fig. 12). Again, the dissolution of limestone which has not been replaced by dolomite produces the present pore geometry. Some of the crinoid ossicles and their optically continuous overgrowths have been replaced by single dolomite crystals, while others presumably remained calcite and have been dissolved to produce vugs. Within these single dolomite crystals, the cloudy original ossicle can easily be differentiated from the clear overgrowth. The production of voids by dissolution of unreplaced patches will be discussed later. Dissolution of the nonreplaced calcite may accompany or follow dolomitization. In other examples where the intermediate stage (partial dolomitization of poorly sorted crinoidal limestones) is observed, the growth of the rhombs is restricted to the carbonate mud matrix. Evidence for mud matrix, and sometimes for mud with abundant floating particles, in those crinoidal sands that are completely dolomitized comes from observation of relict and ghost textures. Pref-

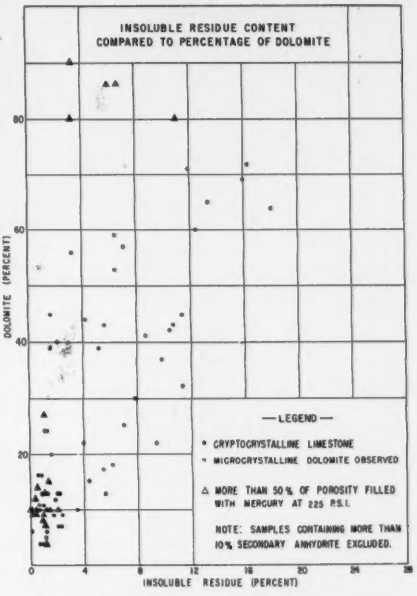
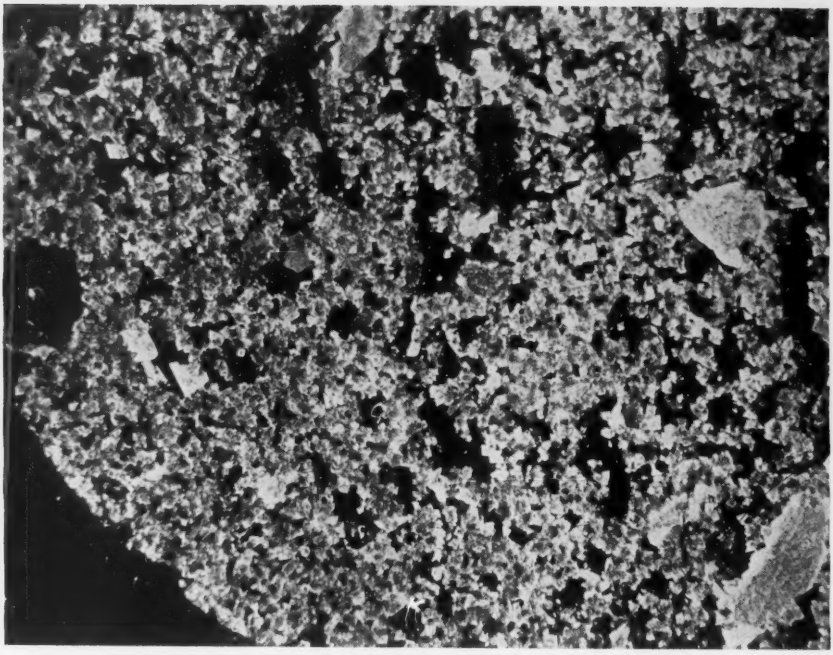


FIG. 11.—Insoluble residue content compared to percentage of dolomite. These samples were taken from the "Midale beds," Charles Formation, Midale field, Saskatchewan. The dolomite was obtained by Versenate analysis, and the residue, largely silt- and clay-size quartz and clay minerals, was determined by weight after removal of anhydrite.

erential growth of dolomite euhedra in mud has been reported by Beales (1958), Bergenback and Terriere (1953), and Hobbs (1957).

It appears that the development of sucrose dolomite results from the growth of randomly oriented, locally uniformly sized replacement dolomite crystals followed by or contemporaneous with the dissolution of the nonreplaced calcite. In many cases, the favored facies is one of quiet-water carbonate mud with contained sand-size particles. However, it is well known that all carbonate muds are not dolomitized, and, conversely, in areas of extensive dolomitization such as the Permian reservoirs of the Central Basin Platform and Northern Shelf, west Texas, carbonate sediments of all degrees of sorting and particle type have been converted to sucrose dolomite.

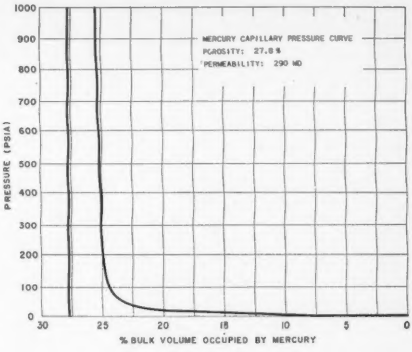
Following or during the formation of su-



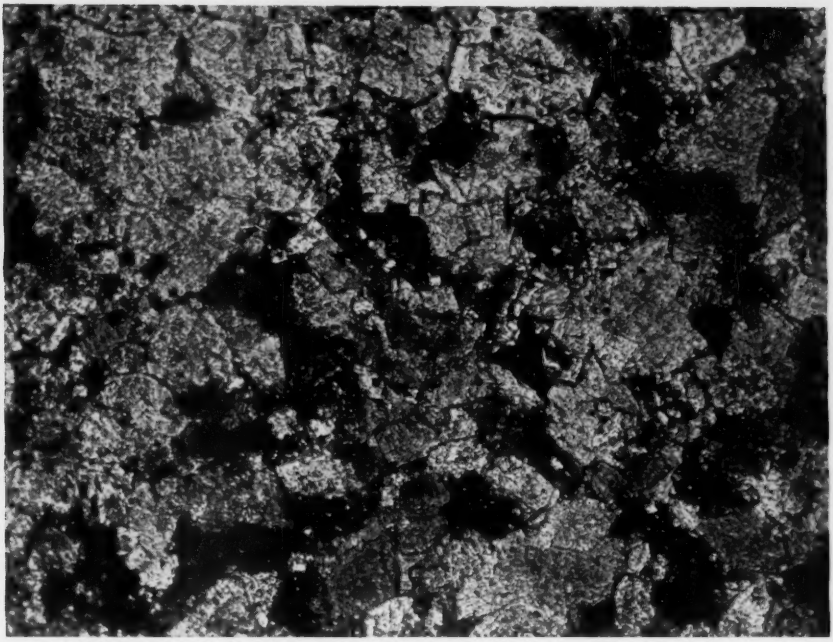
Crossed nicols, ×10.

FIG. 12.—Sucrose dolomite with fossil fragment molds. Comparison with associated rocks, and relict textures indicates that this rock began as a carbonate crinoidal sand with interstitial mud or mud with abundant sand-size particles. Dolomite has grown as euhedra in the mud and has impinged on the crinoid and other fossil fragments. Some of the crinoid fragments and their optically continuous overgrowths have been dolomitized as single crystals with retention of the internal structure of the fragment. Those not dolomitized have been removed (see crinoid columnal mold), as has the other nonreplaced limestone, to produce porosity.

Harmattan field, Alberta; Turner Valley Formation (Mississippian).



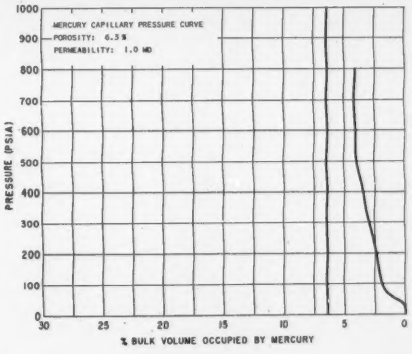
crose dolomite pressure solution and/or simple deposition of optically continuous overgrowth cement may materially reduce porosity and pore size (fig. 13). This is especially critical in the very fine (crystal size less than 20 μ) sucrose dolomites. Here, a small amount of such pore-destructive cement may materially reduce the pore size and thus the permeability, with little change in pore volume. In addition, simple nondolomite cement, such as calcite, anhydrite, or quartz, may reduce porosity and pore size.



Crossed nicols, ×120.

FIG. 13.—Interlocking dolomite. This mosaic of anhedral to subhedral dolomite exhibits little porosity. The dark patches are crystals at extinction and show the irregular outline of interlocking anhedra. This rock may have begun as a sucrose dolomite, similar to figure 8. However, continued optically continuous growth of dolomite has caused the crystals to grow together, filling space and thus reducing porosity and pore size. All gradations between the sucrose texture with its relatively large pore size and the relatively dense interlocking texture are observable, commonly within a single thin section.

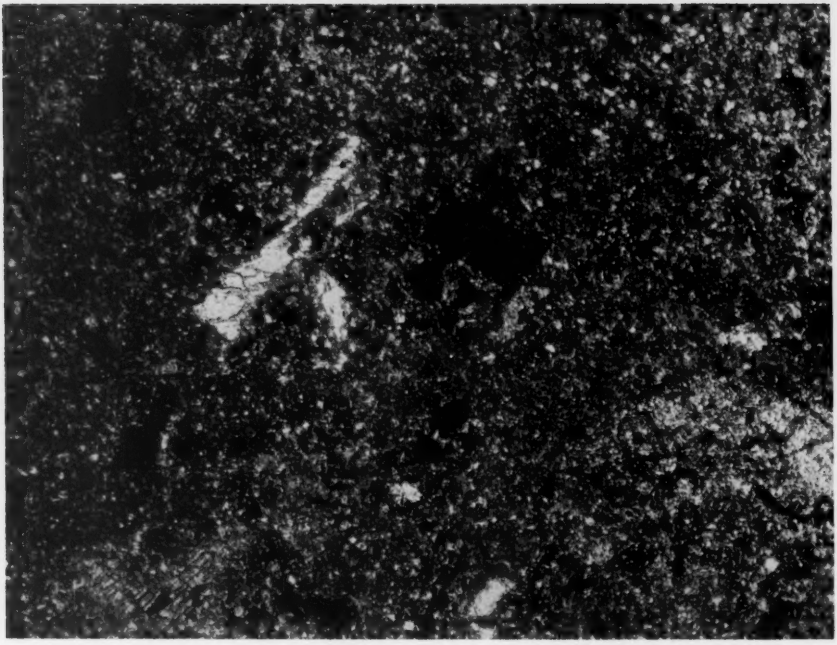
Gas City field, Montana; Red River Formation (Ordovician).



Secondary Vugs

The development of secondary voids larger than the normally expected interparticle voids represents a separate problem in carbonate pore genesis. Several mechanisms are important; some produce an interconnected vug network, and others contribute to pore volume but depend on associated interparticle or intercrystalline pore space for their interconnection.

Anhydrite molds.—Metasomatic anhydrite is commonly observed in subsurface



Crossed nicols, ×50.

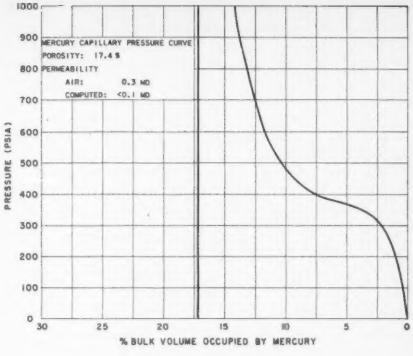
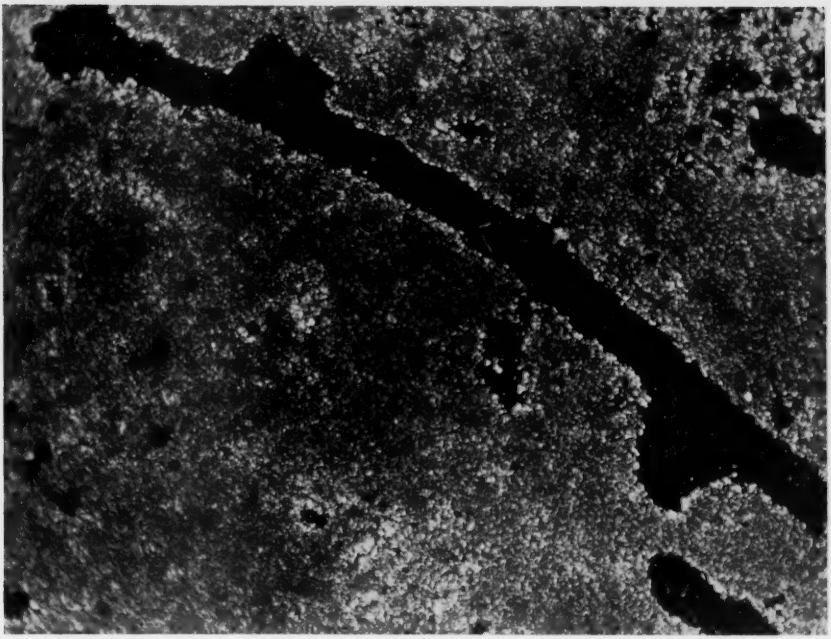


FIG. 14.—Replacement anhydrite molds. Anhydrite molds are formed by the dissolution of replacement anhydrite crystals in a fine sucrose dolomite. Note the rectangular re-entrants and the relatively straight sides of the vugs (black). Anhydrite crystals (white), similar in size and shape to the vugs, are present in the rock. This rock has poor capillary properties despite the 17.4 percent porosity and the large-diameter vugs. This indicates that the vugs are isolated and that they depend on intercrystalline void space for interconnection. The capillary pressure measurements indicate that mercury does not enter these vugs, which comprise approximately 50 percent of the void space, below pressures of approximately 300 psi, or an equivalent pore radius of approximately 0.0003 mm. This lack of interconnection would be expected from vugs produced by the dissolution of isolated replacement anhydrite crystals.

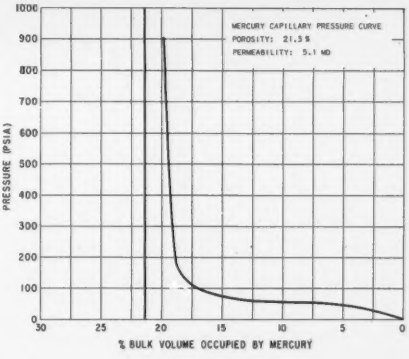
Desert Creek field, Utah; Paradox Formation (Pennsylvanian).



Crossed nicols, ×29.

FIG. 15.—Replacement anydrite molds. This vug is interpreted as having formed by the dissolution of replacement anhydrite. The replacement anhydrite crystal probably began by replacing a fossil shell. Rectangular re-entrants into the fine sucrose dolomite matrix suggest the anhydrite mold origin. Replacement anhydrite is common in these rocks. The capillary pressure curve demonstrates that these vugs are interconnected through the intercrystalline pore space.

Harper field, west Texas; San Andres Formation (Permian).



carbonate rocks, both limestone and dolomite. Gypsum has been observed in rocks from below 3000 ft, but it is exceedingly rare. Anhydrite commonly replaces fossil fragments, but other hosts, such as carbonate mud, nonskeletal particles, or combinations of these, occur in many examples. In many carbonate rocks, simple anhydrite cement growing within void spaces extends into the rock matrix by replacement. Such replacement anhydrite attempts to develop its crystal form, resulting in single crystals with straight sides and rectangular re-entrants and projections. However, the shape of the replacement anhydrite crystal is commonly controlled in part by some element of the rock fabric (fig. 1). Vugs exhibiting straight sides and rectangular re-entrants

and projections are common, particularly in fine sucrose dolomite. These vugs are frequently associated with replacement anhydrite crystals of similar shape and size. In the cases of rocks that contain no other fabric element whose removal would produce rectangular vugs, it has been concluded that the vugs result from dissolution of replacement anhydrite (figs. 14, 15, and 16). Caution must be used in determining that the re-entrants and projections are not formed by euhedral vug lining crystals. These anhydrite molds are common in subsurface carbonate rocks of both Paleozoic and Mesozoic age throughout North America. To the writer's knowledge, specific examples of vugs produced by this mechanism have not previously been reported. Because such vugs are formed by the removal of individual noninterconnected replacement anhydrite crystals, the vugs themselves are dependent on interparticle or intercrystalline porosity for interconnection. Evidence for this is found in very high initial displacement pressures of mercury for rocks with as much as 23 percent porosity, 9 porosity percent of which is formed by anhydrite molds. These high pressures indicate that the vugs are not filled with mercury until the fine intercrystalline pores are entered.

Carbonate dissolution.—Examples of dissolution of selected fabric elements, such as particular fossil fragments (commonly aragonitic) or parts of fossil fragments, have been observed. In addition, selective dissolution of part of the carbonate mud matrix in a poorly sorted crinoidal or other carbonate sand can produce an interconnected vug network resembling interparticle porosity. However, some undissolved mud is commonly observed along with relicts of other crystals, such as quartz and dolomite, that grew in the mud before dissolution. An example of this pore type was reported by Graf and Lamar (1950). They proposed dissolution of interstitial carbonate mud as a porosity-producing mechanism in the Fredonia Oolite.

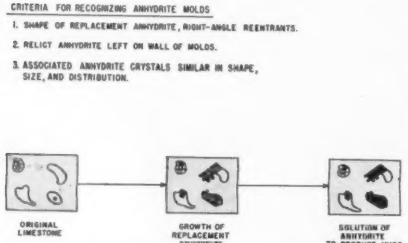
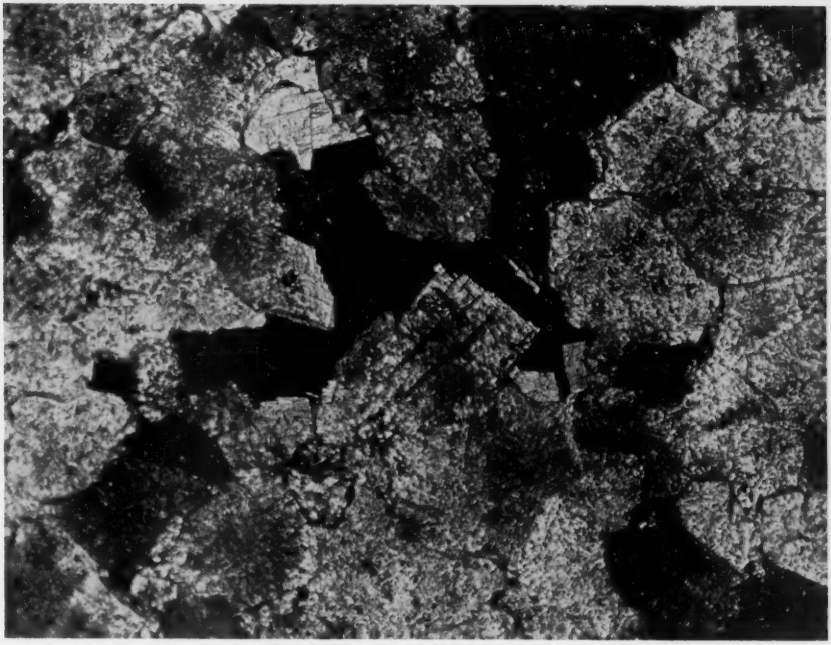

CRITERIA FOR RECOGNIZING ANHYDRITE MOLDS
1. SHAPE OF REPLACEMENT ANHYDRITE, RIGHT-ANGLE REENTRANTS.
2. RELICT ANHYDRITE LEFT ON WALL OF MOLDS.
3. ASSOCIATED ANHYDRITE CRYSTALS SIMILAR IN SHAPE, SIZE, AND DISTRIBUTION.
ORIGINAL LIMESTONE
GROWTH OF REPLACEMENT ANHYDRITE
SOLUTION OF ANHYDRITE TO PRODUCE VUGS

FIG. 16.—Origin of anhydrite mold vugs.

Dolomite.—Distinct from the uniform growth distribution of euhedral replacement dolomite, discussed under Sucrose Dolomite, is patchy or mottled dolomite growth with nonreplaced patches of original calcite. These patches commonly are larger than the replacement crystals, and their dissolution can produce vugs. The texture of such patchy dolomite may be either interlocking and thus space-filling, or mutually interfering with accompanying interdolomite crystal porosity. The resulting pore geometry and porosity will be either wholly vugs with interlocking dolomite texture or vugs plus intercrystalline porosity if mutually interfering. Actually, the mutually interfering texture may be converted to the interlocking texture by pressure solution. The reason for patchy replacement is often not determinable in the final product; however, in the cases of the dolomitized, poorly sorted, crinoidal limestone discussed previously, the monocrystalline crinoid ossicles appear to resist replacement by dolomite euhedra and thus remain calcite. On dissolution, the calcite ossicles and any calcite optically continuous overgrowths that may have developed before dolomitization are removed to produce vugs (fig. 12). Actually, patchy dolomite replacement may be caused by many inhomogeneities in the original limestone—for example, large shell fragments, burrowings, and microfracture networks. Porosity may be reduced or destroyed by simple cementation of these vugs with calcite, dolomite, anhydrite, or quartz (figs. 17 and 18). Such cements are usually clear and may or may not show crystal alignment perpendicular to the vug wall. An example of the former containing two phase hydrocarbon filled fluid inclusions was reported by Murray (1957).

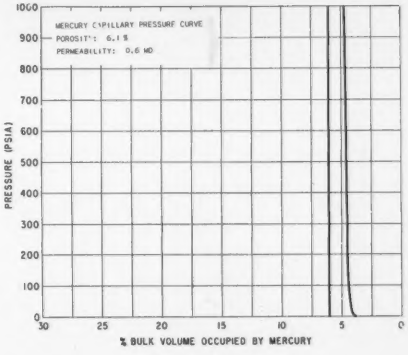
Chert.—A similar textural sequence is observed when limestone is replaced by a lacework of fine chert. Dissolution of the relict limestone has produced reservoir rocks with



Crossed nicols, ×84.

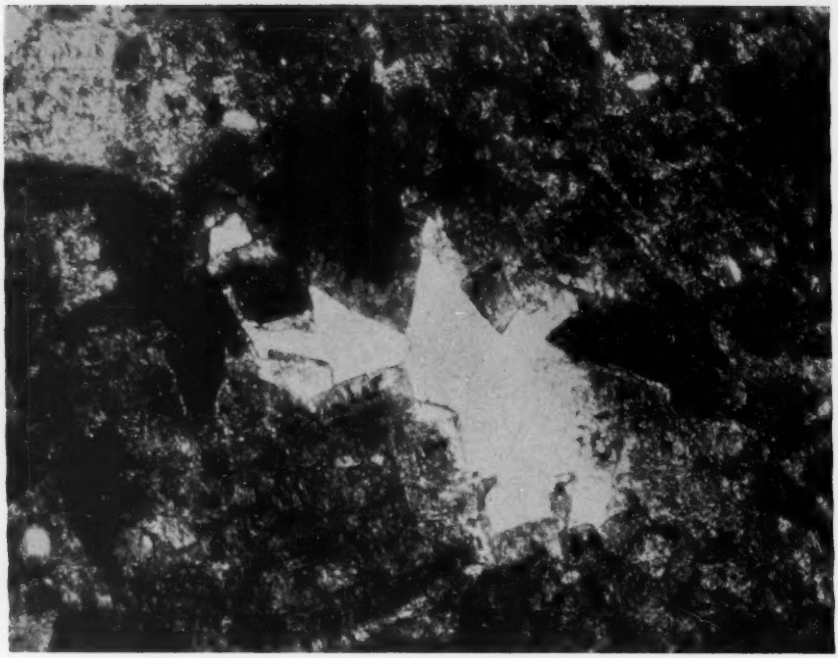
FIG. 17.—Vugs within an interlocking dolomite mosaic. These open, interconnected vugs (see the capillary pressure curve) are found within a mosaic of tightly interlocking dolomite crystals. Some euhedral dolomite has grown out in the vug or has impinged on calcite before dissolution of a nonreplaced calcite patch. Although there is no petrographic evidence for the origin of this vug, it probably represents a nonreplaced patch of calcite that was dissolved or an earlier vug that was not cemented during dolomitization. The air permeability measurement does not agree with the capillary pressure curve. This is to be expected in a low-porosity, vuggy pore geometry. The capillary pressure curve best represents the true size and interconnection of the voids.

Sturgeon Lake field, Alberta; Leduc Formation (Devonian).



as much as 35 percent porosity in the Crossett Devonian field, west Texas, but commonly with small-size pore interconnection because of the fine size of the chert and the completeness of the replacement lacework.

Fractures.—The development of open fractures represents a void-producing mechanism (Waldschmidt and others, 1956). Actually, even in the most intensely fractured subsurface carbonate rock reservoirs, it is doubtful that the total porosity contribution from open fractures will exceed 1 percent; however, the possible contribution of abundant, interconnected open fractures to



Crossed nicols, ×38.

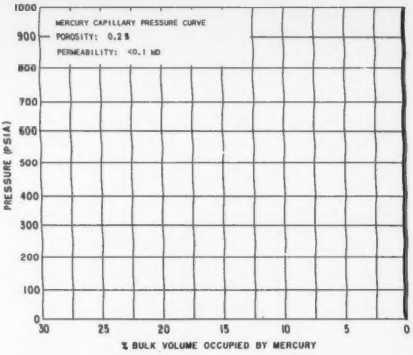


FIG. 18.—Vugs within interlocking dolomite mosaic—cemented with anhydrite. This rock is identical in texture to figure 17, with the exception that the crystal-lined vug has been completely cemented by a single anhydrite crystal. The capillary pressure curve indicates that little, if any, porosity has been retained and demonstrates how tightly the dolomite anhedra are interlocked. This cementation is associated with thin lenses of bedded anhydrite.

Sturgeon Lake field, Alberta; Leduc Formation (Devonian).

fluid flow should be recognized. The enlargement of fractures by dissolution of carbonate host rock is sometimes observed. Fractures, like other voids, are subject to simple cementation, and this is commonly observed. Indeed, it is not uncommon to find both filled and unfilled fractures in the same rock.

Void preservation.—Vugs in either interlocking or mutually interfering dolomite may result from dolomitization of a limestone with an original vug network or interparticle porosity. If the pore space is not filled with dolomite cement and if the dolomite crystal size is sufficiently smaller than

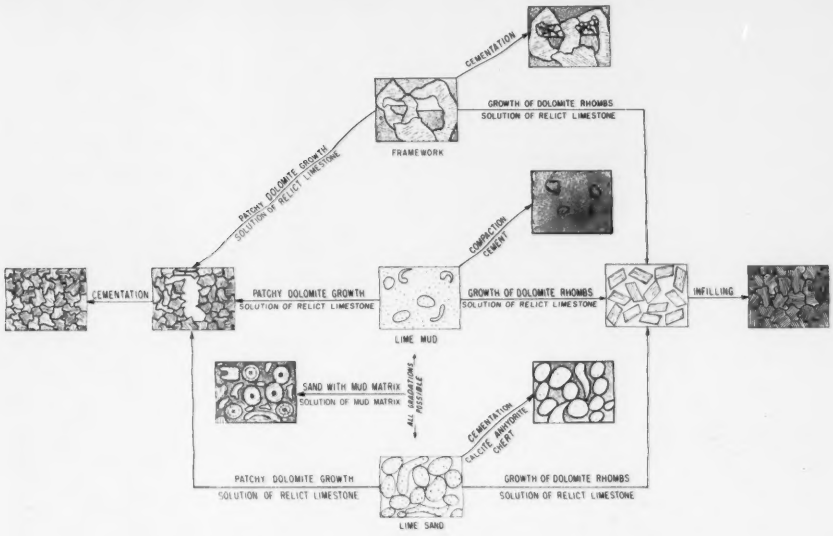


FIG. 19.—Diagram illustrating the evolution of some pore types.

the original void size, these voids may be preserved. This pore type is sometimes observed in fine sucrose dolomite where the dolomite appears to have replaced only the particles and to have preserved an original interparticle porosity. However, it is usually impossible to determine whether the original interparticle space had been filled with sparry cement which resisted dolomite replacement and which was later dissolved with the nonreplaced parts of the particles.

SUMMARY

Figure 19 illustrates diagrammatically some of the mechanisms discussed in this paper. The original sediments—framework, mud, and sand—are depicted in the central column. Diagenetic processes changing porosity and pore-size distribution appear to the right and left. These processes and others discussed in the paper appear to be significant in making or destroying carbonate porosity in North American hydrocarbon reservoirs. However, as void space receives more petrographic attention, other processes will undoubtedly be recognized, and a better understanding of those already known will become available. The latter will undoubtedly be aided by a better understanding of the physico-chemical processes that operate to produce the observed textures and textural changes.

ACKNOWLEDGMENT

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 30, NO. 1, PP. 85–90
FIG. 1, MARCH, 1960

POROSITY THROUGH DOLOMITIZATION: CONSERVATION-OF-MASS REQUIREMENTS[1]

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ABSTRACT

Although the process of dolomitization is not well known at present, the requirement of conservation of mass of the chemical constituents of the rock and the interstitial solution places definite limits on the process. Most subsurface waters contain relatively small concentrations of bicarbonate compared to their magnesium and calcium contents. For these waters, the process of dolomitization must proceed by a mole-for-mole exchange of calcium by magnesium without macrotransport of carbonate. This can be expressed by the chemical equation

$$Mg^{2+}+2CaCO_3 \rightarrow MgCa(CO_3)_2+Ca^{2+}.$$

The porosity of the rock is not only affected by the 13 percent volume shrinkage on mole-for-mole dolomitization, but it also depends on the amount of compaction which takes place. If a lime mud is dolomitized, the mud will be compacted as the dolomite crystals grow, resulting in a decrease in porosity. Once the dolomite crystals form a space-supporting framework, the rate of compaction will decrease considerably. Replacement of the remaining mud will then lead to a marked increase in permeability and may also result in a porosity increase.

INTRODUCTION

In the preceding paper, Murray (1960) has investigated the petrographic evidence for the relationship between porosity and dolomitization. At the present time, the process of dolomitization is not well understood. It is impossible to predict how much dolomite would be deposited if water of a given composition traversed a particular piece of rock under specified conditions of temperature, pressure, and temperature gradient. Before such predictions can be made, more will have to be known about the solution chemistry of carbonate rocks.

From discussions with Murray, it became evident that a number of the relationships observed by him could be understood in terms of basic principles and did not require a complete understanding of the process of dolomitization. It is the purpose of this paper to state these principles explicitly and to show how they can be used to describe the history of dolomitization of a hypothetical lime mud. Only by combining observations of the geologic record with theory can we gain a better understanding of the mechanisms that alter sedimentary rocks. It is hoped that such a combination has been achieved by the preceding paper and this one and that a better basis has been provided for understanding the process of porosity development through dolomitization.

[1] Shell Development Company Publication No. 207. Manuscript received June 19, 1959.

CONSERVATION-OF-MASS RELATIONS

Dolomitization is the process whereby limestone becomes dolomite by the substitution of magnesium carbonate for a portion of the original calcium carbonate (A.G.I. Glossary). The only generally abundant minerals of calcium and magnesium in the subsurface are calcite, dolomite, and anhydrite. Their properties are listed in table 1.

Let us assume that a volume of rock contains only these three minerals and that the pore space is filled with an interstitial solution. At some initial time ($t=0$), there are $x(0)$, $y(0)$, and $z(0)$ units of mass of calcite, dolomite, and anhydrite, respectively, per unit volume of the rock. The fractional porosity $\phi(0)$ will be

$$\phi(0)=1-\frac{x(0)}{\rho_c}-\frac{y(0)}{\rho_d}-\frac{z(0)}{\rho_a}, \qquad (1)$$

where ρ_c, ρ_d, and ρ_a are the densities of calcite, dolomite, and anhydrite, respectively. As dolomitization takes place, the mineral

TABLE 1.—*Properties of calcite, dolomite, and anhydrite*

	Calcite	Dolomite	Anhydrite
Formula	$CaCO_3$	$CaCO_3 \cdot MgCO_3$	$CaSO_4$
Molecular weight	100.09	184.42	136.15
Density, gm/cc	2.711	2.872	2.96
Molar volume, cc/gm mole	36.9	64.2	46.0
Mass fraction Ca	0.400	0.217	0.294
Mg	0.000	0.132	0.000
CO_3	0.600	0.651	0.000
SO_4	0.000	0.000	0.706
	1.000	1.000	1.000

composition of the rock will change. In addition, compaction may result, and, if so, the original unit volume of rock will experience a fractional compaction ζ, resulting in a rock volume $V(t)$ of $1-\zeta$. At this time the porosity is

$$\phi(t)=1-\frac{\frac{x(t)}{\rho_c}+\frac{y(t)}{\rho_d}+\frac{z(t)}{\rho_a}}{1-\zeta}. \qquad (2)$$

The change in mineral composition is accomplished by a transport in solution of the constituents of the minerals. These constituents are calcium ion, magnesium ion, carbonate ion, and the sulfate ion. Since the carbonate may be transported as bicarbonate or as dissolved carbon dioxide, we must consider the change in total carbonate, regardless of the form in which it is transported. The change in mass of the four constituents in the element of rock is related to the change in mass of the three mineral phases by the following four equations:

$$\Delta Ca=0.400[x(t)-x(0)]+0.217[y(t)-y(0)]+0.294[z(t)-z(0)], \qquad (3)$$

$$\Delta Mg=0.132[y(t)-y(0)], \qquad (4)$$

$$\Delta CO_3=0.600[x(t)-x(0)]+0.651[y(t)-y(0)], \qquad (5)$$

$$\Delta SO_4=0.706[z(t)-z(0)]. \qquad (6)$$

Since equations (3) through (6) are four linear equations involving only the three minerals, only three of the constituents can be varied independently.

In general, the amount of the constituents in the interstitial solution is negligible compared to the amount in the mineral phases. For instance, in a dolomite with 40 percent porosity whose interstices are filled with sea water, the ratio of the mass of the constituent in the rock to the mass in the solution will be 2300, 400, and 20,000 for calcium, magnesium, and carbonate, respectively. A change in composition of the stationary interstitial solution can thus produce only a negligible change in the mineral composition. To produce a significant change, the constituents must be transported into the element of rock in solution, either by diffusion or by fluid flow.

RELATIVE CONCENTRATIONS OF CALCIUM ION, MAGNESIUM ION, AND CARBON DIOXIDE IN SUBSURFACE WATERS

Most subsurface brines contain a large amount of calcium and magnesium and a relatively small amount of CO_2 (by CO_2 is meant all three forms, CO_2, HCO_3^-, and CO_3^{2-}). For instance, of 923 reported analyses of brines from oil-productive formations in Oklahoma (Wright and others, 1957), 800 contained more than ten times as much magnesium by weight as bicarbonate. The calcium content was even higher than that of magnesium. About 50 had at least five times as much magnesium as bicarbonate, and only 42 contained more bicarbonate than magnesium.

To understand why the CO_2 content of ground waters is generally low, we must consider the mechanisms that affect it. The atmosphere contains only a very small

amount of CO_2 (0.03 volume percent). When equilibrium is reached with the atmosphere, the CO_2 content of rain water, or waters containing neutral salts, is very low. The CO_2 content can be raised above this value by four mechanisms.

Waters containing alkaline salts can pick up appreciable amounts of CO_2 from the atmosphere (for example, sodium bicarbonate waters).

Thermal decomposition of carbonate rocks due to igneous intrusions can increase the CO_2 content of the water. This process is limited only by the amount of CO_2 that can go into solution at the particular temperature and pressure. Excess CO_2 above this will form a separate gas phase.

Dissolution of carbonates will cause an increase in CO_2. In the absence of other acidic materials, the solubility of carbonates depends primarily on the initial amount of CO_2 in the water. At saturation, essentially a solution of the bicarbonate of the mineral is formed; thus, in a water containing carbonic acid as the only acidic material, solution of carbonates would approximately double the CO_2 content of the water.

Oxidation of organic carbon will lead to an increase of CO_2 in the water. In order for oxidation to take place, oxygen must, of course, be supplied, and any CO_2 produced will be able to leave the water by the same route by which the oxygen is supplied. Thus, oxidation is not likely to be a frequent mechanism by which the CO_2 content is significantly affected.

DOLOMITIZATION BY WATERS CONTAINING RELATIVELY SMALL AMOUNTS OF TOTAL CARBONATE

As mentioned above, it is very common for ground waters to contain relatively small concentrations of total carbonate compared to the concentrations of calcium and magnesium ions. At the present time, we do not know enough about the solution chemistry of calcite and dolomite to predict the process of dolomitization by moving ground waters. In the case of low relative concentrations of total carbonate, however, it is possible to show that dolomitization must take place approximately by the mole-for-mole replacement of calcite without introduction of external CO_2. This can be represented by the equation

$$Mg^{2+}+2CaCO_3 \rightarrow CaMg(CO_3)_2+Ca^{2+}.$$

In this equation, only the ions are transported over macroscopic distances by the moving ground waters.

First, we can base the argument on an economical use of the ground water in dolomitization. As a typical example, let us consider the following brine (No. 660 of Wright and others, 1957):

	ppm	moles/1000 gm Solution
Calcium	11,300	0.283
Magnesium	2,330	0.096
Bicarbonate	37	0.0006

During the process of dolomitization, if the magnesium concentration decreases by one percent, 0.001 mole of dolomite can be produced per 1000 gm of solution flowing through the rock by a mole-for-mole magnesium-calcium exchange without affecting the bicarbonate concentration. The change in magnesium concentration is about twice the total amount of bicarbonate present in the water. A one-percent change in the bicarbonate could lead to a precipitation of only 0.000006 mole of magnesium carbonate or to a dissolution of a like amount of calcite. Therefore, for a given percentage change in the composition of the solution, about 200 times as much dolomite could be produced by a mole-for-mole replacement with the local rock used as a source for carbonate as could be produced if magnesium carbonate were brought in.

A second argument can be based on a consideration of the calcite equilibrium. In a previous paper (Weyl, 1958) the author has shown that under geologic conditions the water in the interstices of a limestone is always in solution equilibrium with calcite. Let us assume that for 1000 gm of the water in the previous example, 0.0002 mole of dolomite is precipitated by the reaction

$$Mg^{2+}+Ca^{2+}+2CO_3^{2-} \rightarrow CaMg(CO_3)_2.$$

For the concentrations in the water, most of the CO_2 is in the bicarbonate form. The water must therefore supply 0.0004 mole of carbonate by the reaction

$$HCO_3^- \rightarrow H^+ + CO_3^{2-}.$$

The decrease in HCO_3^- and increase of H^+ will disturb the equilibrium of calcite:

$$CaCO_3+H^+\rightleftarrows Ca^{2+}+HCO_3^-.$$

The concentration product constant for this reaction is

$$K_c=\frac{[Ca^{2+}][HCO_3^-]}{[H^+]}.$$

Before 0.0002 mole of dolomite has precipitated, if the hydrogen ion concentration is assumed to be 10^{-7} ($pH\approx 7$), the concentration product constant has a value of

$$K_c=\frac{0.283\times 6\times 10^{-4}}{10^{-7}}=1698.$$

After the dolomite has precipitated, if no solution of calcite is assumed, the product will be

$$K_c'=\frac{0.283\times 2\times 10^{-4}}{4\times 10^{-4}}=0.1415.$$

If we assume that this slight change in salinity of the solution has no effect on the activity coefficients, the product must remain the same in order that the solution remain saturated with respect to calcite. To accomplish this, x moles of calcite must be dissolved per 1000 gm of solution to give a concentration product constant

$$K_c=1698=\frac{(0.283-x)(2\times 10^{-4}+x)}{(4\times 10^{-4}-x)}.$$

Solving for x, we find

$$x=3.999\times 10^{-4}\approx 4\times 10^{-4}.$$

In order for the solution to remain saturated with respect to calcite, therefore, approximately 2 moles of calcite must be dissolved for every mole of dolomite precipitated. This is equivalent to stating that dolomitization takes place by a mole-for-mole replacement of magnesium for calcium, where the only carbonate used is that locally available by dissolution of calcite.

For no carbonate exchange, equation (5) becomes

$$\Delta CO_3=0=0.600[x(t)-x(0)]+0.651[y(t)-y(0)]; \quad (7)$$

therefore,

$$\Delta x=-1.085\Delta y. \quad (8)$$

It should be pointed out that the removal of calcite is not restricted to the immediate space where a dolomite crystal is growing. If no CO_2 is introduced, equation (8) must be satisfied by dissolution of calcite in the general area of dolomite precipitation. Half the calcium and all the carbonate so dissolved will diffuse to the growing dolomite crystal. The assumption of no net carbonate transport is valid only over volumes of the order of a cubic inch. On a microscale, transport by diffusion to the site of crystal growth does take place. This, however, will not affect the average porosity.

Dolomitization without macrotransport of carbonate will, of course, take place only as long as the total carbonate content of the water is small compared to the concentrations of calcium and magnesium. As the relative amount of total carbonate is increased, the above relationship will become less valid. In view of the prevalence of low-carbonate waters in rocks, the above mode of dolomitization is of extreme geological importance. In applying these ideas to particular geological examples, one must be careful that the carbonate content of the water in the pores of the rock is and always was low compared to the concentrations of calcium and magnesium.

MOLE-FOR-MOLE DOLOMITIZATION OF A LIME MUD

In order to determine how the above-derived principles can be applied to a particular situation, we shall consider the hypothetical mole-for-mole dolomitization of a lime mud. We shall start with 1 cc of lime mud having a porosity before dolomitization of 40 percent. The volume of lime mud will thus contain 1.627 gm of calcite. As magnesium ions are brought in, calcite will dissolve, and a number of dolomite crystals will begin to grow. During this replacement, the relationship between volume of calcite removed and volume of dolomite precipitated is given by

$$\Delta V(\text{calcite})=-\Delta V(\text{dolomite})\times 1.15. \quad (9)$$

If dolomitization proceeds without compaction, the porosity of the mud must increase, since the volume of calcite dissolved exceeds the volume of dolomite precipitated

by 13 percent. It is very unlikely, however, that part of the lime mud will dissolve without compaction. A more reasonable assumption is that the porosity of the mud part of the system will remain the same or will even decrease, owing to the weight of the overburden. Compaction of the mud will continue until the larger crystals of precipitated dolomite form a space-supporting framework. Once this takes place, the overburden will essentially be supported by the dolomite crystals, and compaction of the lime mud should be materially reduced. This shift from dolomite crystals floating in the mud to a dolomite framework containing some mud will be gradual and will occur during the time in which the dolomite changes from 40 percent to 70 percent of the volume of the rock.

In order to make a quantitative estimate, let us assume that until the dolomite forms a supporting framework, the mud retains a porosity of 40 percent through compaction. When the dolomite occupies 55 percent of the rock, we shall assume that it forms a space-supporting framework and that compaction ceases.

We further assume that the magnesium ion is being introduced at a constant rate of 0.132 gm during a time interval t^*. Since the mass fraction of Mg^{2+} in dolomite is 0.132, this means that 1 gm of dolomite will be precipitated in the initial cubic centimeter of mud in time t^*. The change in mass of calcite in this element is then given by equation (8), and we obtain

$$\text{Mass of calcite} \quad x(t)=1.627-1.085\frac{t}{t^*}, \tag{10}$$

$$\text{Mass of dolomite} \quad y(t)=1.000\frac{t}{t^*}. \tag{11}$$

The volume of calcite is

$$\frac{x}{\rho_c}=0.600-0.400\frac{t}{t^*}, \tag{12}$$

and the volume of dolomite is

$$\frac{y}{\rho_d}=0.348\frac{t}{t^*}. \tag{13}$$

Until the dolomite forms a supporting framework, the lime mud has a porosity of 40 percent, so the total volume, which is equal to one minus the fractional compaction ζ, is

$$\begin{aligned}1-\zeta&=1-(0.667-0.348)\frac{t}{t^*}\\&=1-0.319\frac{t}{t^*}\quad|\ t<t'.\end{aligned} \tag{14}$$

The porosity, ϕ, is (see equation 2)

$$\phi(t)=1-\frac{0.600-0.052\dfrac{t}{t^*}}{1-0.319\dfrac{t}{t^*}}\quad|\ t<t'. \tag{15}$$

The dolomite comprises 55 percent of the volume of the rock at a time t'; therefore,

$$0.348\frac{t'}{t^*}=0.55\times\left[1-0.319\frac{t'}{t^*}\right],$$

or

$$t'=1.052t^*. \tag{16}$$

At this time, the volume has been reduced to 0.664 cc because of compaction, and the porosity will be 0.179. Beyond this point, little further compaction should take place, since the dolomite is now assumed to form a space-supporting framework. The porosity is, therefore,

$$\phi(t)=1-\frac{0.600-0.052\dfrac{t}{t^*}}{0.664}\quad|\ t>t'. \tag{17}$$

The dolomitization process will be completed when all the calcite has been dissolved. This will happen when $t=1.500t^*$. At this time, the porosity is 0.214. The final part of the dolomitization process is thus accompanied by an increase in porosity, owing to the fact that the volume of calcite dissolved exceeds the volume of dolomite precipitated.

The fractional dolomitization, which is the weight of dolomite divided by the weight of calcite plus dolomite, is

$$\frac{y}{x+y}=\frac{\dfrac{t}{t^*}}{1.627-0.085\dfrac{t}{t^*}}.$$

Figure 1 is a plot of porosity as a function of fractional dolomitization.

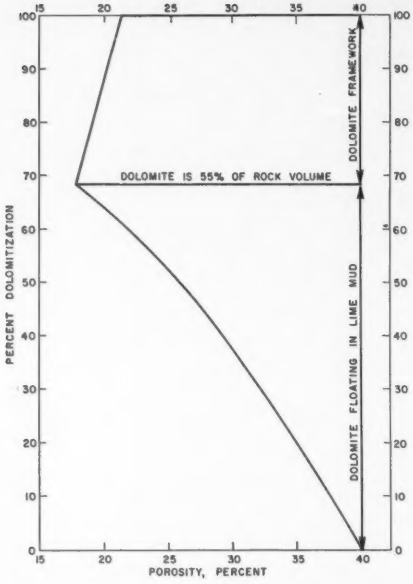

FIG. 1.—Porosity versus percent dolomitization.

During the early part of dolomitization, the permeability of the mud will decrease slightly, since the dolomite occupies space previously occupied by mud. After the dolomite forms a space-supporting framework, replacement of the mud in the spaces between the dolomite crystals will result in a marked increase in permeability. The final permeability will be considerably greater than that of the lime mud, owing to the larger size of the dolomite crystals. Roughly, the permeability of the final sucrose dolomite will exceed the permeability of the mud by a factor equal to the square of the average diameter of the dolomite crystals divided by the square of the average diameter of the lime mud particles and crystals.

CONCLUSIONS

In waters containing relatively low concentrations of total CO_2, dolomitization will take place by mole-for-mole replacement of calcium by magnesium without introduction of carbonate. Initially, dolomitization of a lime mud will tend to reduce the porosity and permeability of the rock. Once the dolomite crystals form a space-supporting framework, the rate of compaction will be reduced, owing to the relatively large size of the dolomite crystals relative to the size of the lime mud particles. Dolomitization beyond this point will lead to an increase in porosity and to a marked increase in permeability.

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FIGS. 1–23, MARCH, 1960

SOURCES OF RECENT SEDIMENTS IN THE NORTHERN GULF OF MEXICO[1]

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ABSTRACT

Sand-size sediments are contributed to the northern Gulf of Mexico by a large number of streams and, to a lesser extent, by marine erosion of coastal deposits, mainly of Pleistocene age. The heavy mineral assemblages and distributive provinces of most sources can be recognized in the Holocene sediments of the northern Gulf of Mexico. Five heavy mineral provinces have been established, the Eastern Gulf, Mississippi, Western Gulf, Texas Coast, and Rio Grande provinces. The Western Gulf province is of complex origin and consists of sediments deposited on the continental shelf during the early Holocene rise in sea level by the Mississippi, Rio Grande, Colorado, Brazos, and Trinity-Neches Rivers. The changing pattern of alluvial plains, deltas, shore deposits, and several major changes in the direction of longshore drift and currents during the transgression can be recognized.

The Texas Coast province consists of a mixture of material contributed by the rivers of the area and of detritus eroded from the Pleistocene deposits of the area by waves. Longshore drift to the west is an agent in the distribution of sediments derived from the Colorado River in the zone between Galveston Bay and Aransas Pass. This material enters the bays from the Gulf side. Barrier island and bay sands are of local origin west of Aransas Pass and consist of a mixture of material derived from the Pleistocene on the land side and material brought in from the shelf during and after the transgression.

INTRODUCTION

The sediments of the northern Gulf of Mexico have been derived from a vast, almost continent-wide, source area and from source rocks ranging in age from Precambrian to Quaternary. They have been weathered under a variety of climates and have been eroded from their source beds by glacial, fluviatile, and marine agents. The sediments are brought to the sea by two large river systems, the Mississippi and the Rio Grande, and by numerous smaller streams. In addition, some material has been derived by marine erosion from the Pleistocene deposits in the coastal area.

In this complex situation an investigation of the sources and transportation of sediments can be expected to contribute much to the understanding of the sedimentation and history of the basin. In addition, it may furnish information concerning the basic laws of sediment supply and transportation in basins with vast and complex hinterlands.

More specifically, the objectives of the present study can be defined as follows:

1.—to define the areas in which sediments directly derived from the Rio Grande and Mississippi are being deposited, and to study the mixing of this material with detritus from other sources;
2.—to identify the sources of sediments on the continental shelf between the distributive provinces of the Mississippi and the Rio Grande, and to study the mechanics and sequence of their deposition; and
3.—to determine the sources of sediments in the bays of the Texas Coast. These sediments may have been derived from local Pleistocene deposits, supplied by local rivers or by longshore drift from the east or west, or brought in from the continental shelf.

Earlier investigations have shown that distribution patterns are not necessarily the same for different size fractions in a single basin and that different sources and different transportation paths may account for the deposition of gravels, sands, and clays (see, for example, van Andel and Postma 1954, p. 155–157). *The present paper deals exclusively with the sources of sand and of the sand fraction of the other sediments. Its conclusions do not necessarily apply to gravels and clays.*

[1] The investigation was supported by a grant from the American Petroleum Institute, Project 51, at the Scripps Institution of Oceanography, University of California, La Jolla. Manuscript received July 22, 1959.

The gravels in this area are of very minor importance but information concerning the source of clays would be of great value.

The study is based on mineralogical analysis. A clear understanding of the problem of sources and transportation requires additional information, for example, morphological and hydrographical observations and studies of the grain size distribution of the sediments. Some of this information is available in various publications on the sedimentology of the basin (Shepard, 1956; Curray, in press; Rusnak, in press) and will be utilized here.

Traditionally, the mineralogical composition of sands has been studied after separation into a light and a heavy fraction. In order to obtain a quantitative picture of the relative contributions of all sediments, not only the compositions but also the relative proportions of both fractions should be known. The present study is restricted to the heavy minerals only because this method offers a rapid technique for the investigation of large numbers of samples and because a sizeable body of information concerning the heavy mineral distributions of other modern depositional basins is available for comparison. Curray (in press) has made a limited study of the light fraction, which agrees reasonably well with the conclusions of the present investigation.

Restriction to the heavy mineral composition imposes limitations upon the conclusions that can be drawn. Studies of both light and heavy fractions of river sands have shown that different sources may possess great differences in the ratio light/heavy minerals (Koldewijn, 1955). Consequently, the effect of a contribution from a certain source in a mixed sediment can be either exaggerated or minimized depending on the relative percentages of heavy minerals in the source materials. In order to obtain reliable estimates of the relative amounts contributed by multiple sediment sources, a study of the percentages of heavy minerals per unit sediment in each of them is required. For the Gulf of Mexico only a small amount of this information is available.

Various factors modify the mineralogical composition of a sediment between source rock and depositional site. In the present study the most important are the elimination of unstable components by weathering and selective sorting of minerals by size and density. In conclusions concerning sediment sources in the Gulf of Mexico such factors will have to be considered.

The heavy minerals of Gulf Coast sediments have been the subject of various earlier investigations. The results of these investigations will be utilized at various places in this study, and ample use had been made of published analyses. The investigation is based upon approximately 800 new heavy mineral analyses from the Mississippi delta, the continental shelf, the coastal zone of Texas, and the rivers draining into the Gulf. A composite Upper Pleistocene section from several offshore wells south of Grand Isle, Louisiana, has been made available by courtesy of the Humble Oil and Refining Company. A sample location map is presented in figure 1. In this paper a selection of characteristic analyses is included for illustration purposes. A complete copy of all data with detailed location maps has been filed in the Library of the Scripps Institution of Oceanography at La Jolla, California.

METHODS OF ANALYSIS

The methods of analysis used in this investigation have been described in detail in various publications (Doeglas, 1940; van Andel, 1950; Poole, 1958, for heavy mineral analysis; Poole, 1957, for grain size analysis by means of a settling tube). The analyses include only the non-opaque minerals. Small percentage variations and the absence or presence of low frequency minerals have no significance because of the restriction to 100 grains per count (Dryden, 1931; Muller, 1943; van Andel, 1950, 1955). To facilitate comparison with data obtained in the present study analyses published by Dohm (1936), Russell (1937), Hough (1937), Goldstein (1942), and Bullard (1942) have been converted on the basis: sum of non-opaques = 100 percent. Only those analyses that contained more than 100 non-opaque grains were used for the conversion.

The size fraction used here is 0.500 to 0.062 mm. The higher minimum size as compared to standard procedure (0.030 mm approximately) reduces the percentage of zircon. This mineral has little diagnostic value and tends to become concentrated in

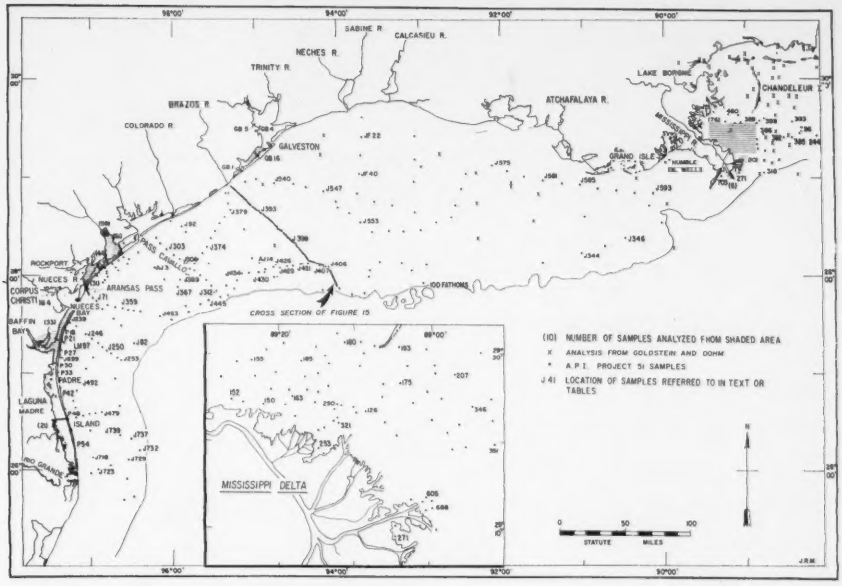

FIG. 1.—Map of place names and sample locations. Numbers indicate samples used in text and tables. See also figure 2 (river samples) and 21 (Rockport area).

fine-grained samples. Table 1 illustrates the difference between analyses of the standard size fraction and analyses of the fraction used here.

SEDIMENT SOURCES

The source areas of the sediments in the northern Gulf of Mexico can be discussed conveniently under four headings: (1) the

TABLE 1.—*Comparison of heavy mineral assemblages obtained by analysis of the 0.500 to 0.030 mm and 0.500 to 0.062 mm grades, respectively, in sands of Texas and Louisiana rivers, in percent of total non-opaque fraction*

	Location at bridge in highway	Size Grade	Hornblende[1]	Tourmaline	Epidote	Zircon	Garnet	Staurolite	Kyanite	Titanite	Rutile	Basaltic Hbl.	Sillimanite	Augite	Hypersthene
Brazos R.	U. S. 79	0.50–0.06		24	5	23	36	4	2	2	4				
		0.50–0.03		15	9	33	32	2		1	8				
Calcasieu R.	U. S. 190	0.50–0.06		22	6	28	3	12	19		9		1		
		0.50–0.03		14	5	61	2	1	5		12				
Colorado R.	U. S. 77	0.50–0.06	82	1	1	2	9	2		2	1				
		0.50–0.03	69	2	7	13	5		1	1	2				
Sabine R.	La. 12	0.50–0.06	8	24	25	12	3	10	13	1	3		1		
		0.50–0.03		7	9	53	3	3	6		19				
Guadelupe R.	U. S. 59	0.50–0.06	2	17	14	33	18	3	4	3	4		2		
		0.50–0.03		6	8	59	6	1	4	5	11				
Trinity R.	U. S. 79	0.50–0.06	2	34	8	19	6	9	11	4	5	1			1
		0.50–0.03		6	12	57	6	5	1		13				

[1] Not including basaltic hornblende.

Mississippi basin, (2) the Rio Grande basin, (3) the other rivers of Texas and western Louisiana, and (4) the rivers east of the Mississippi delta. The following brief discussion of the four groups has been based largely on a study of the literature. Some sediments, particularly in the bays of Texas, appear to have been derived to a large extent from local Pleistocene and early Holocene[2] deposits. The evidence for this conclusion will be discussed in connection with the description of the specific depositional areas.

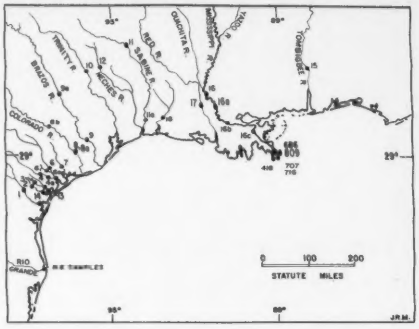

FIG. 2.—Location of river samples listed in tables 2 and 3.

Mississippi River Basin

The heavy minerals of the Mississippi and its major tributaries have been studied in detail by Hough (1937) and Russell (1937). The typical assemblage found along the entire course below the confluency with the Des Moines River consists of augite and hornblende with appreciable percentages of epidote and garnet (table 2). The assemblage is very similar to the suites described by various authors for glacial and outwash deposits in the northern United States and Canada (Doeglas, 1949; Dreimanis and others, 1957; Gravenor, 1951; Kruger, 1937; Murray, 1953). Russell and Hough conclude that at the present time the Mississippi carries to the Gulf of Mexico material essentially derived from glacial deposits. This conclusion may require some modification because the contributions of non-glacial material by the Red, Ouachita, Arkansas, and Yazoo Rivers may be under-represented in the heavy mineral assemblage as a result of a low heavy mineral content of their sands. Nevertheless, it seems plausible to assume that the bulk of the Mississippi sediment deposited in the Gulf of Mexico is of glacial origin.

[2] Throughout this paper the term "Holocene" (synonymous with but preferred to Recent) will be used to designate the series of sediments deposited on top of the last marine unconformity in the Gulf of Mexico.

Rio Grande Basin

The Rio Grande contributes to the Gulf of Mexico an assemblage of heavy minerals

TABLE 2.—*Heavy mineral composition of selected Mississippi River sands in percent of total non-opaque fraction*[1]

Sample number		Hornblende[2]	Tourmaline	Epidote	Zircon	Garnet	Staurolite	Kyanite	Titanite	Basaltic Hbl.	Augite	Hypersthene
16	Mississippi River	34	6	20	3	16	1		3	3	11	3
16a	Mississippi River	37	1	17	2	7			2	1	32	1
16b	Mississippi River	40	3	16	3	6			3	3	25	1
16c	Mississippi River	32	1	21	3	7			3	1	28	4
716	South Pass	33	5	14	2	13	1		5	1	25	1
707	South Pass	41		12	2	5			2	4	34	
478	Southwest Pass	36	3	16	7	9	1		2	2	22	2
686	Pass a Loutre	38	3	14		8			3	5	28	1
809	Pass a Loutre	30	1	22	3	13		1	4	3	21	2
17	Atchafalaya R.	28		16		32			4	2	18	

[1] For sample locations see fig. 2.
[2] Not including basaltic hornblende.

TABLE 3.—*Heavy mineral composition of sands from Texas and Louisiana rivers,[1] in percent of total non-opaque fraction.*

Sample number		Hornblende[2]	Tourmaline	Epidote	Zircon	Garnet	Staurolite	Kyanite	Titanite	Basaltic Hbl.	Augite	Hypersthene	Rutile, Sillim.
RG	1	21	3	25	3	12	1	2	2	12	16		3
	7	31		19	2	5		5	2	8	25	1	2
	9	19	2	25	15	12	1	2	4	5	12		3
	13	17	3	25	8	5		2	4	10	24	2	
	16	26	2	15	3	4	1	4		10	35		
	17	21		15	3	4		1	4	15	35		2
1		4	15	14	20	15	5	3	5	2	11		6
2			25	14	33	20		3	3				2
3		1	28	14	30	18	4	1					4
4		6	8	12	40	12	2	13	2				5
4a			14	13	45	14	1	8	1				4
5			13	5	51	13	6	2	2				8
6		2	17	14	33	18	3	4	3				6
6a		1	13	16	56		9	3	1				1
7		2	5	34	39	14	1	3		1			1
8		80	1	4	6	5			4				
8a		69	2	7	11	4			4				3
8b		82	1	1	2	9	2		2				1
9			24	5	23	36	4	2	2				4
9a		1	18	9	22	37	8						5
10		2	34	8	19	6	9	11	4	1		1	5
11		1	18	13	23		21	15	1				8
11a		8	24	25	12	3	10	13	1				3
12			19	7	29	2	11	15	2				15
13			10	8	66	3	8	1					4
14		1	10	16	49	12	4	3	1				4
18			22	6	28	3	12	19					10
15			12	21	8	8	18	23	2				8

[1] For sample locations see fig. 2.
[2] Not including basaltic hornblende,

which in many respects is similar to that of the Mississippi (table 3). In both rivers there are considerable variations between individual analyses, and single samples cannot always be differentiated. In larger groups the Rio Grande assemblage is found to contain more pyroxene and less hornblende than the Mississippi suite. It is also characterized by small but persistent percentages of basaltic hornblende (probably derived from young volcanics in northeastern Mexico, Sidwell and Renfroe, 1943) and metamorphic minerals, mainly kyanite. Another difference which will be discussed later is in the varietal composition of the hornblendes.

Information is scarce concerning the assemblages of the Rio Grande above the delta and of its tributaries. Rittenhouse (1944) has decribed a suite from the middle course of the river which is quite similar to the assemblage found in the delta. However, the middle course sediments contain less epidote and garnet and a higher amount of hornblende. Addition of Pecos material which contains an abundance of epidote and garnet (Sidwell, 1944) probably accounts for the higher percentages of these minerals in Rio Grande delta sediments.

Rivers of Texas and Western Louisiana

The drainage basins of the rivers between the Mississippi and Rio Grande consist primarily of sedimentary formations of Cretaceous, Tertiary, and Quaternary age. Some Paleozoic is exposed in the head-

waters of the Colorado and Brazos rivers. The Nueces and Colorado Rivers also drain relatively small areas of igneous and metamorphic rocks.

The heavy mineral associations (table 3, fig. 2) of most rivers in this area reflect the mineral composition of the Cretaceous and Quaternary as described by Cogen (1940; see also Bornhauser, 1940; Rogers and Dawson, 1958; Rogers and Powell, 1958; Sidwell, 1937, 1947). The influence of Tertiary material is less important. The principal association consists of zircon and tourmaline with minor amounts of epidote, garnet, staurolite, and kyanite. Individual differences between rivers exist, but their significance cannot be inferred from the limited material studied. Higher percentages of kyanite and staurolite are found in the Trinity, Neches, Sabine, and Calcasieu Rivers which drain extensive areas of Paleocene and Eocene sediments which are characterized by a staurolite-kyanite association (Cogen, 1940; Sun, 1954).

The zircon-tourmaline association of the Nueces River contains small but significant percentages of augite and basaltic hornblende which give this assemblage an intermediate position between that of the Rio Grande and those of most other Texas rivers. These minerals may have been derived from Tertiary volcanics in the Uvalde region in the headwaters of this river.

The Colorado River association is completely different from all others in the Texas coastal area. It is characterized by an almost pure blue-green hornblende suite which is found as far upstream as the Llano uplift but is absent above this area of pre-Cambrian rocks (Sidwell and Cole, 1937). The hornblende assemblage delivered to the Gulf of Mexico has been derived either directly from this Precambrian or indirectly from Pleistocene terraces in the Colorado valley (Mathis, 1944).

Area East of the Mississippi River

The heavy mineral suites that characterize shelf sediments in the area east of the Mississippi delta consist of kyanite and staurolite with some zircon and epidote. The very limited information available in the literature (Needham, 1934; Blankenship, 1956; Potter, 1955) and one sample from the Tombigbee River (fig. 2, table 3) suggest that the source of this assemblage is the Cretaceous and younger mantle of the southern Appalachians rather than the old core which, according to the few publications available (for example, Carroll, 1957; Carroll, Neuman, and Jeffe, 1947), would only supply a zircon assemblage.

Summary of Sediment Sources

The heavy mineral assemblages supplied by rivers to the Gulf of Mexico can be divided into several groups, each characterized by a well-defined heavy mineral assemblage. In table 4 and figure 3 the composition and sources of sand-size sediments have been summarized. In addition, some material is derived by marine erosion from Pleistocene deposits of the Texas coastal region.

HEAVY MINERAL PROVINCES OF THE NORTHERN GULF OF MEXICO

Following Edelman (1933) a *sedimentary petrological province* is defined as "a group of sediments which constitute a natural unit by age, origin and distribution." Each province is characterized by a light and a heavy mineral association. Transition zones of mixing, overlap and interfingering normally separate adjacent provinces. Within the boundaries of each province the composition of the individual samples may vary considerably as a result of variations in grain size, sampling and analytical errors, statistical variation, or incomplete mixing of sediment from multiple sources. Consequently, the associations have to be defined in general semiquantitative terms. A province is best defined when its association contains minerals that do not occur in significant amounts in any of the other provinces of the same basin. An example is the Eastern Gulf province (table 5). Frequently, however, the same minerals are found in many or all of the provinces of a sedimentary basin, and associations have to be defined with the aid of differences in frequency. Under such circumstances the identification of single samples may be difficult, and the establishing of province boundaries and the evaluation of mixed zones become to a certain extent a subjective matter. Since most of the provinces of the northern Gulf of Mexico have to be defined on the basis of quantitative

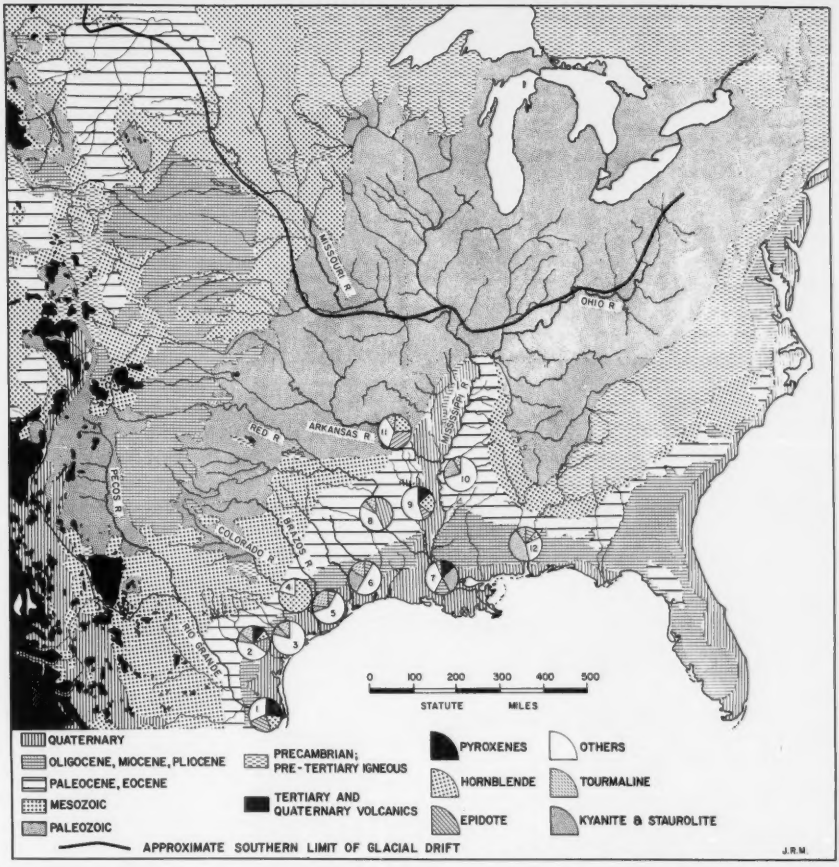

FIG. 3.—Regional geology of northern Gulf of Mexico source area, and heavy mineral composition of sediments supplied by principal rivers. Geology simplified from "Geological Map of North America," Geol. Soc. America (1946). Mineral assemblages: 1-Rio Grande, 2-Nueces River, 3-composite of streams between Nueces and Colorado, 4-Colorado River, 5-Brazos River, 6-composite of Trinity, Sabine and Neches Rivers, 7-Mississippi River, 8-Red River, 9-Ouachita River, 10-Yazoo River, 11-Arkansas River, 12-Tombigbee River.

differences, the study of sediment sources, transportation, and mixing is particularly difficult. Therefore, an attempt has been made to improve the resolving power of the heavy mineral method by introduction on a large scale of the use of single mineral parameters and mineral ratios. The following five heavy mineral provinces can be distinguished in the northern Gulf of Mexico (fig. 4; table 5). Four of them were defined originally by Goldstein (1942).

Eastern Gulf province: characterized by a kyanite-staurolite association derived from the Cretaceous and younger sedimentary mantle of the Appalachians.

Mississippi province: characterized by a hornblende-pyroxene association with appreciable amounts of epidote and

TABLE 4.—*Composition and source of heavy mineral assemblages supplied to Northern Gulf of Mexico*[1] *in percent of total non-opaque fraction. Averages of all available samples from the lower courses of rivers*

Rivers	No. of samples	Sediment Source	Hornblende[2]	Tourmaline	Epidote	Zircon	Garnet	Staurolite	Kyanite	Basaltic Hbl.	Pyroxenes	Others
Mississippi	16	Glacial of northern U. S.	28	5	16	5	21	1	1	5	18	
Rio Grande	7	Upper Rio Grande, Pecos; volcanics of Chihuahua	20	2	22	5	6	1	3	10	26	4
Nueces	3	Cret., Tert., Quat., and volcanics of Uvalde	4	15	14	20	15	5	3	2	11	11
Rivers between Neuces and Colorado	10	Cret., Tert., and Quat. of Central Texas	1	14	15	44	13	4	4			5
Colorado	5	Llano Uplift, pre-Cambr.	77	1	4	7	6	1				4
Brazos	3	Cret., Tert., and Quat. of Central Texas	1	21	7	22	36	6	1			6
Trinity, Neches, Sabine, Calcasieu	5	Tert. and Quat. of East Texas and West Louisiana	1	17	10	37	4	11	12			8
Area east of Mississippi River	1	Cret., Tert., and Quat. mantle of S. Appalachians		12	21	16	8	18	23			2

[1] For sample locations see fig. 3.
[2] Not including basaltic hornblende.

garnet. The assemblage has been derived from glacial deposits.

Western Gulf province: characterized by a hornblende association with some epidote but little or no pyroxene, garnet, or basaltic hornblende. The origin of this province is very complex and will be discussed below.

Rio Grande province: characterized by a pyroxene-hornblende association with associated epidote and significant percentages of basaltic hornblende. The sediments are supplied by the Rio Grande.

Texas Coast province: characterized by a tourmaline association with abundant zircon and some epidote. Texas rivers, but not the Colorado and Rio Grande, and Pleistocene deposits of the Texas coast supply the material. In many samples small amounts of hornblende are present, which do not belong to the tourmaline association but are the result of admixture of material from the Western Gulf province.

Within the Western Gulf province a group of samples can be distinguished by means of their high zircon content ("Brazos Group"). This group is a mixture of Western Gulf material with another association probably derived from the Brazos and Trinity Rivers. This association has not been observed in a pure form on the shelf.

Except for the Western Gulf province the sources of all sediments in the northern Gulf of Mexico can be established with certainty. The Western Gulf association resembles rather closely the assemblage supplied by the Colorado River, thus suggesting the rather surprising conclusion that this relatively minor river would have supplied sediment for a very large portion of the continental shelf. Goldstein (1942), who first observed the similarity, has offered an alternate explanation for the provenance of the Western Gulf sediments by suggesting that they may have been derived from Mississippi material that had lost its unstable pyroxenes by selective weathering. It will be shown later that the Western Gulf province is a complex system with multiple sources including Colorado sediments and weathered Mississippi and Rio Grande material.

TABLE 5.—*Average heavy mineral composition of associations characterizing Northern Gulf of Mexico provinces, in percent of total non-opaque fraction*

Province	Source	Hornblende[2]	Tourmaline	Epidote	Zircon	Garnet	Staurolite	Kyanite	Pyroxenes	Basaltic Hbl.	Others	No. of samples
Eastern Gulf	Cret., Tert., Quat. S. Appalachians	13	12	16	12	2	16	16	3	t	10	31
Mississippi	Mississippi River	40	2	16	2	9	t	t	25	2	3	116
Western Gulf	complex	58	5	17	4	7	1	1	3	1	3	127
Rio Grande	Rio Grande River	23	3	15	6	10	1	2	24	7	9	41
Texas Coast	Rivers between Nueces and Brazos; reworked Pleistocene	7	28	11	23	9	8	5	3	t	6	54
Brazos Group[1]		32	8	15	20	8	3	3	3	1	7	19

[1] No pure samples available.
[2] Not including basaltic hornblende.

SELECTIVE SORTING OF HEAVY MINERALS

Heavy minerals cover a broad range of densities and sizes. Many minerals, for example zircon, have a marked preference for one specific size. Fine-grained sediments are almost invariably much richer in zircon than coarser deposits from the same source. Many other minerals are concentrated in specific size fractions but the range depends on the source. Consequently, sorting by size can result in strong modification of the mineral assemblage. With a suitable set of preferred sizes a hornblende-pyroxene-epidote suite can be separated into a coarse pyroxene-hornblende and a fine hornblende-epidote assemblage (van Andel, 1955).

This example calls to mind the difference between the pyroxene-hornblende associations of the Rio Grande and Mississippi, and the hornblende association of the Western Gulf province. A test is required to establish whether the difference between these associations is caused entirely or partly by sorting instead of provenance.

In figures 5 and 6 percentages of all diagnostically important minerals have been plotted against the median diameter of the sand fraction. The heavy minerals are compared with the sand fraction only because a light fraction with a lower size limit of 0.10 mm is approximately hydraulically equivalent to the heavy fraction which has a lower limit of 0.062 mm.

The graphs show that except for hornblende none of the mineral species exhibit a correlation with the median of the grain size distribution. Apparently, the percentages of augite, kyanite, epidote, and tourmaline are not controlled by sorting. The hornblende percentages increase slightly with decreasing size. This applies only when samples from all provinces are combined and appears to be due largely to the fact that those provinces that are low in hornblende consist of somewhat coarser sediment than those high in hornblende. The grain size ranges of the individual associations overlap to such an extent that the systematic differences in hornblende content that exist between them cannot be explained by sorting alone.

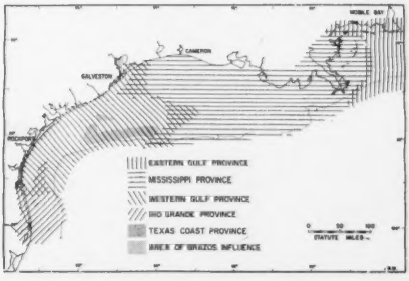

FIG. 4.—Heavy mineral provinces in the northern Gulf of Mexico. Sample locations figure 1. For details concerning Mississippi Delta, Laguna Madre, and the bays of Rockport, Texas, see figures 7, 18 and 22 respectively.

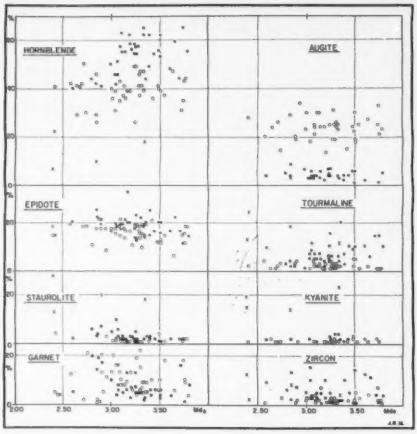

FIG. 5.—Comparison between mineral percentage and median of size distribution of sand fraction in samples from the Mississippi (open circles), Western Gulf (black dots), and Eastern Gulf (crosses) provinces.

Within each individual association grain size and hornblende percentage are not at all or not significantly co-variant. Minor differences in hornblende content may be caused by sorting, but in general it can be concluded that grain size does not control the mineralogical composition.

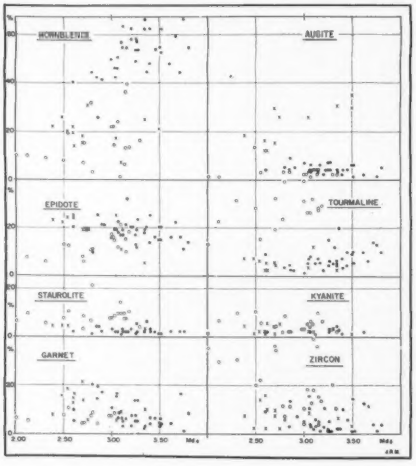

FIG. 6.—Comparison between mineral percentage and median of size distribution of the sand fraction in samples from the Western Gulf (black dots), Texas Coast (open circles), and Rio Grande (crosses) provinces.

HEAVY MINERAL DISTRIBUTION, MISSISSIPPI DELTA REGION

The heavy mineral distribution in the Mississippi delta and on the adjacent shelf (fig. 7, table 6) is the result of the interplay of sediments from two sources, the Mississippi River and the Cretaceous and younger mantle of the southern Appalachians. At present, direct supply of Appalachian sediments is probably restricted to the river estuaries and the nearshore zone east of the Mississippi delta. On the shelf transportation of material from this source appears to be restricted to reworking and redistribution of sediments already present.

From the Mississippi delta the hornblende-pyroxene association spreads east in a broad arc a little beyond the Chandeleur Islands and grades through a narrow transition zone into the Appalachian kyanite-staurolite association. The dividing line runs approximately north and south, cutting obliquely across a major tongue of modern Mississippi sediment extending eastward from the active Pass a Loutre subdelta (Scruton, 1955; Shepard, 1956). The distribution of Mississippi material as inferred by those authors from lithological and petrographical observations does not agree with the heavy mineral distribution (fig. 8). Shepard has explained this discrepancy by the assumption that periodically Eastern Gulf sand containing kyanite and staurolite assemblages is moved westward over the shelf and mixed with Mississippi material supplied from the east.

The heavy mineral boundary of figure 8 is based on single mineral percentages. A similar boundary has been established by Goldstein (1942) on the basis of heavy mineral associations. A parameter more sensitive to changes in relative contribution of Mississippi and Eastern Gulf material is the ratio hornblende/kyanite+staurolite. In the pure associations of the two sources these minerals are mutually exclusive. Contours based on this ratio (fig. 7) reflect better the mixing of Mississippi and Eastern Gulf material than the association map and conform much better to the shape of the Mississippi tongue shown in figure 8. Apparently, Mississippi

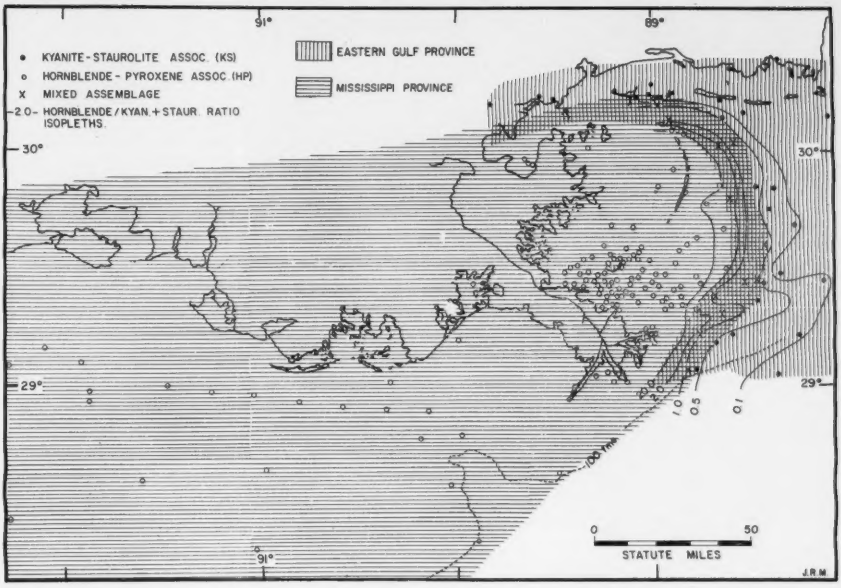

FIG. 7.—Distribution of heavy mineral assemblages, eastern Mississippi Delta area. From 173 samples a total of 96 percent conform to pattern. Samples from Goldstein (1942) included after conversion.

minerals are moved much farther east within the tongue than to either side, although a strong dilution with Eastern Gulf material remains undeniable.

The sediments of the Mississippi tongue consist of almost pure clay with a very small percentage of fine sand. Consequently, the heavy mineral content is low (less than 0.15 percent by weight). The sandy deposits of the Eastern Gulf province on the other hand, contain an average of 0.40 percent heavy minerals (Goldstein, 1942). Thus, even a minor contribution of Eastern Gulf material would greatly change the heavy mineral composition of the sediments of the Mississippi tongue. It can be estimated that in that part of the tongue where the hornblende/kyanite+staurolite ratio is 1.0 the contribution of Eastern Gulf material does not exceed 5 percent.

The Mississippi province overlaps considerably in an eastward direction on deposits of the Eastern Gulf province. Sediments belonging to the Eastern Gulf province have been found in various places under the surface of the Mississippi delta (Shepard, 1956; Shepard and Lankford, 1959), but the precise western limit of the province is not known as far as the author is aware.

HEAVY MINERAL DISTRIBUTION WESTERN GULF PROVINCE

The Western Gulf province has been defined by Goldstein (1942) on the basis o a

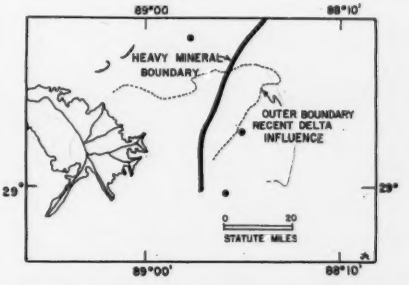

FIG. 8.—Relation between modern Mississippi sediment tongue and heavy mineral province boundary in the area east of the Mississippi Delta. Simplified after Shepard (1956, fig. 38).

TABLE 6.—*Representative analyses of samples from the Mississippi and Eastern Gulf Provinces; in percent of total non-opaque fraction*

Sample number	Hornblende[1]	Tourmaline	Epidote	Zircon	Garnet	Staurolite	Kyanite	Titanite	Basaltic Hbl.	Augite	Hypersthene	Others
Mississippi Province												
MP 126	44	1	18	3	4	2			3	23		2
BS 152	46	1	14	5	2	1		4	2	25		
155	47	7	16	2	6	1		1	2	17	1	
163	50	2	15	2	6	2		2		18	1	2
175	34	3	23		8			3	1	26	2	
180	27	3	14	3	17			2		32		2
185	42	3	12	1	10			6	4	18	3	1
BI 193	41	4	17	2	5		1	4	2	24		
BC 253	43	5	17	1	6			1		27		
BB 271	47		13	2	10		1	5	3	16	1	2
MP 290	48	1	13		3	1		3	5	24		2
321	46	3	14	3	12		2	1	2	13	3	1
PL 346	39	1	15	3	16	1		2	1	18	1	3
351	38		25	2	7			4	4	19		1
GM 366	37	2	19		8			3		29	1	1
398	37	2	21	2	15	1	1	3	2	16		
399	43	3	14	2	8	1		3	2	22	2	
BS 460	49		13	6	5			1	4	18		4
PL 605	44	2	10		10	1		3	1	26	2	1
688	41	1	17		3			5	2	29	1	1
SP 703	39		9	1	4			3		41		3
J 344	44	2	20	4	10			1		18		1
346	49	1	10	5	6				2	26	1	
575	29	2	15	2	16	1		4	1	29		1
581	26	1	19	3	19	1	1	4		22	3	1
585	47	3	9	1	7		1	4	1	26		1
593	31		13	3	21			6		25		1
Eastern Gulf Province												
GM 244	11	11	13	11	2	19	20	2		4		7
318	11	9	20	23	4	7	12	2		2		10
385	10	19	19	8	1	20	14			4		6
386	7	12	18	12		28	15					8
392	18	11	14	4	5	18	23	1		4		1
393	22	24	14	2	4	13	19					2

[1] Not including basaltic hornblende.

relatively small number of samples. Its hornblende association is characterized by high hornblende percentages and less than 5 percent pyroxene (table 7). The large number of samples available for the present study demonstrates that this province is not a homogeneous unit and that it probably consists of more than one association. A separate group of samples can be distinguished (Brazos Group) which is characterized by a zircon-tourmaline assemblage with some kyanite and staurolite. This group contains highly variable percentages of hornblende, suggesting that all samples contain admixtures of material from the hornblende association.

Theoretically, all rivers emptying in the northwestern Gulf of Mexico could contribute sediment to the Western Gulf province or have done so in the past. The deposits on the shelf may have been derived either directly from the rivers or indirectly from

TABLE 7.—*Representative analyses of samples from the Western Gulf Province, in percent of total non-opaque fraction*[1]

Sample number	Hornblende[2]	Tourmaline	Epidote	Zircon	Garnet	Staurolite	Kyanite	Titanite	Basaltic Hbl.	Augite	Hypersthene	Others
Western Gulf Province												
J 71	66	4	19	2	3				3	3		
82	50	2	17	13	5	2	2		7	2		
92	66	8	12	2	4		1	1	1	4	1	
239	38	12	8	15	12	1	1	3			1	9
246	52	2	19	1	10	3		1	1	6	2	3
250	48	6	19	1	9			1	5	6		5
253	53	5	30		2			1	2	5		2
303	55	5	29	2	5	1			1	1	1	
306	38	5	25	5	10	3	1	7	2	1	2	1
309	42	4	20	7	5	1	7	2	2	4	3	3
310	43	6	15	6	11	3	7	4	1	1		3
312	59	8	4	4	6	2	2	2	3	7		3
359	59	3	19	1	2	1	1	1	3	8	1	1
367	55	10	16	1	8		4	2		3	1	
369	48	5	13	10	9	3	1	3	1	3		3
374	32	6	24	11	13	5	1	2	1	2	1	2
379	55	6	20	1	5	2		1	1	5		4
393	64	4	22		1	1		1	1	4	1	1
399	32	6	27	9	12	4	3	2		1	1	3
406	36	4	18	14	7		2	1	1	13	1	3
445	53	10	11	3	10	3	1	1	1	7		
453	56	3	15	4	6	1	3		5	5	1	1
492	58	5	17		3	1	1	3	4	7	1	
540	58	4	20	3	7			3	2	2		1
547	42	5	19	10	16			1	1	3	1	2
553	41	2	20	5	8	1	2	2		17	2	
699	30	13	16	8	16	3	7	2		1	1	3
AJ 3	54	6	23	1	5	2	2		3	3	1	
14	40	9	20	9	5	2	2	2	2	2	2	5
JF 22	49	4	13	7	5	2	1	1	2	15		1
40	44	2	5	7	17	1	1	3		18	2	
"Brazos Group"												
J 407	24	7	13	32	7	2	4	3	2	2		4
421	40	10	15	9	8	4	4	1	2	4		3
425	21	11	9	34	9	5	2	2	2	4		1
426	14	14	20	29	8	3	4		1	2	1	4
430	11	15	20	23	8	3	6	6			1	7
434	7	5	12	54	5	4	4	1	1	1		5

[1] For sample locations see fig. 1.
[2] Not including basaltic hornblende.

Pleistocene or early Holocene sediments accumulated on the Gulf Coastal Plain or on the shelf. Prior to the post-Pleistocene transgression many of the older deposits have been subjected to substantial weathering on land and on the exposed continental shelf. As a result less stable components may have been removed. Goldstein (1942) has suggested that removal in this manner of the pyroxenes of the Mississippi association may have produced the material for the hornblende association. Similarly, one might assume that selective removal of the pyroxenes from the Rio Grande association

TABLE 8.—*Average composition of Mississippi and Rio Grande suites after removal by computation of all pyroxene in excess of 5 percent, compared with average composition of Western Gulf suite, in percent of total non-opaque fraction. Data from table 5.*

	Hornblende[1]	Tourmaline	Epidote	Zircon	Garnet	Staurolite	Kyanite	Pyroxenes	Basaltic Hbl.	Others
Mississippi	52	2	21	2	11			5	3	4
Rio Grande	29	4	18	8	12	1	3	5	9	11
Western Gulf	58	5	17	4	7	1	1	3	3	3

[1] Not including basaltic hornblende.

would result in an assemblage similar to the hornblende association.

On the average the Western Gulf assemblage contains a maximum of 5 percent pyroxene. In table 8 are listed the assemblages that result when all pyroxene in excess of 5 percent is removed from the Mississippi and Rio Grande associations by computation. The Rio Grande association thus artificially weathered is not similar to the hornblende association; it contains too little hornblende. The Mississippi suite, on the other hand, can be transformed into a suite quite similar to the Western Gulf assemblage.

That such weathering of the Mississippi association actually has occurred with results identical to those predicted by calculation is demonstrated by the heavy mineral sequence of a composite section of the late Quaternary off Grande Isle, Louisiana (fig. 9). The assemblages in the upper part of the section, though variable, are similar to the modern Mississippi association. At the top of the Prairie Formation (Upper Pleistocene), however, a marked and sudden decrease of pyroxene and a concomitant increase of hornblende occur. Samples from the Prairie and the underlying Montgomery Formations have assemblages quite similar to those of the hornblende association. The decrease of pyroxene is accompanied by a strong increase in the degree of etching and alteration of this mineral and an increase of the number of deeply altered minerals (alterites), suggesting that selective weathering is the cause of the disappearance of the pyroxene rather than a change in source. The abundance of fragile hacksaw structures indicates that the weathering took place in place. Thus, weathered Mississippi sediments appear to be a potential source of material for the Western Gulf province. It may be noted that the hornblendes of the Prairie and Montgomery Formations show no signs of weathering.

Both the weathered Mississippi and the Western Gulf suite are quite similar to the Colorado association. Weathered Rio Grande material, though different in a pure form, would be difficult to recognize in mixtures with one of the three others. All of these assemblages contain hornblende as a principal component.

They differ greatly, however, in the vari-

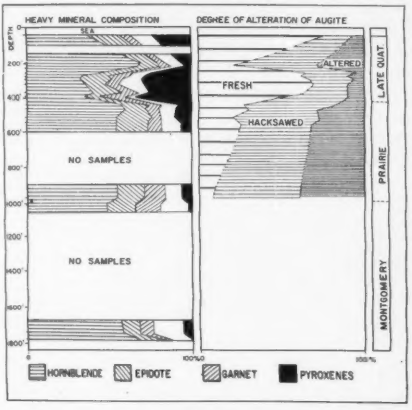

FIG. 9.—Heavy mineral variation in a composite section of the Upper Pleistocene in Humble Oil Company wells, Grand Isle, Block 18, Louisiana. Location of section in figure 1.

TABLE 9.—*Average composition of hornblende fraction in Mississippi, Rio Grande, and Colorado sediments, in percent of total hornblende. Averages of five samples for each river*

	Blue-green	Common	Brown	Basaltic	Others	Ratio blue-green to brown plus common
Mississippi	53	22	16	6	3	1.4
Colorado	89	7	4			8.0
Rio Grande	26	34	14	1	25	0.5

etal composition of the hornblende fraction. The sediments of the northern Gulf of Mexico contain four main types of hornblende: *blue-green* (metamorphic) *hornblende* with strong pleochroism from blue-green to green; *common hornblende*, moderate to weak pleochroism from green and brown to brownish green; *brown hornblende*, weak pleochroism in shades of brown, often deeply colored; and *basaltic hornblende*. In addition small percentages of other amphiboles (tremolite, actinolite) have been found. Percentages of hornblende varieties based on a count of 100 hornblende grains have been obtained for sets of representative samples from the Mississippi, Rio Grande, and Colorado rivers (table 9) and for all Western Gulf samples. In view of the difficulty of identification, in particular of the brown and common varieties, the percentages are approximations only. This may in part explain why the separate percentages of brown and common hornblende do not appear to have diagnostic value. The ratio of blue-green to common plus brown hornblende (called *hornblende ratio* in the following discussion), on the other hand, is a parameter of great value. Basaltic hornblende percentages in general are too low to be of much value.

With the aid of the hornblende ratio it is possible to distinguish between material derived from the Rio Grande, Mississippi, and Colorado Rivers respectively, even where it occurs in minor quantities. The other potential sources of sediment, the Brazos, Neches, Trinity, Sabine, and all other minor Texas rivers, can be distinguished by means of characteristic minerals (zircon, tourmaline, kyanite, and staurolite) which do not occur in significant quantities in the Mississippi, Rio Grande, and Colorado. The percentages of these minerals and the hornblende ratio provide a set of parameters which can be used to identify the relative contributions of the various sources in different parts of the Western Gulf province. These parameters and their characteristic values for all sources are listed in table 10, and the areal distributions are presented in the maps of figure 10–13. Presentation of individual values is impossible on small scale maps, but an estimate of the reliability of the patterns can be obtained with the aid of the percentages of samples that conform to the pattern shown on the map. For every map this criterion is indicated.

Essentially, the patterns have been constructed on the basis of the parameter values alone. In some cases, however, the data allow of two or more slightly different interpretations. In such cases the morphological and grain size observations published by Curray (in press) have been utilized to select the most plausible pattern. In evaluat-

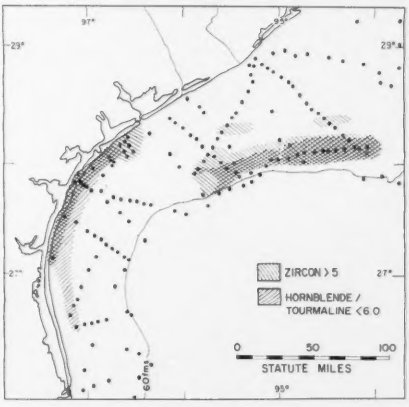

FIG. 10.—Areal distribution of zircon percentages and hornblende/tourmaline ratios in the Western Gulf province; 97 percent of zircon percentages and 99 percent of hornblende/tourmaline ratios conform to pattern.

TABLE 10.—*Mineral parameters characterizing sediments of all potential sources of supply for the Western Gulf Province*

Source	Hornblende[1] %	Pyroxene %	Zircon %	Staurolite plus Kyanite %	Hornblende ratio (see text)	Ratio Hornblende[1] to Tourmaline
Mississippi	40	25	<5	<1	1–2	>20
Mississippi, weathered	52	<5				
Rio Grande	23	24	<5	<5	<1	1–10
Rio Grande, weathered	29	<5				
Colorado	77		<5		>5	>20
Rivers east of Brazos	1		>20	5–10		<1
Brazos, Pleistocene and rivers west of Colorado	1–10		30–40	>10		<1

[1] Not including basaltic hornblende.

ing the maps it must be borne in mind that in some areas the parameter shown pertains to a major portion of the sand-sized sediment but in others the parameter pertains only to a small fraction. In the area close to the shelf edge south of Galveston, for example, the hornblende ratio map (fig. 12) shows a predominance of Colorado derived sediment. Comparison with the hornblende percentage map (fig. 13), however, demonstrates that the sand fraction of this area contains very little of this mineral and that the bulk of the sand must have had other sources. Hence, for a proper evaluation of sediment sources, all maps must be used in combination.

In general the maps show reasonable agreement. The discrepancies that can be observed locally are, in part at least, caused by the relatively high error inherent in an analysis based on counts of 100 grains per sample. The influence of the statistical counting error is particularly great in mineral ratios but has been reduced by omitting all ratios where one or both components occur with less than 5 percent.

The information that can be obtained from the mineral parameter maps has been summarized in figure 14. This sketch map shows the regional distribution of all pure and mixed assemblages in the Western Gulf

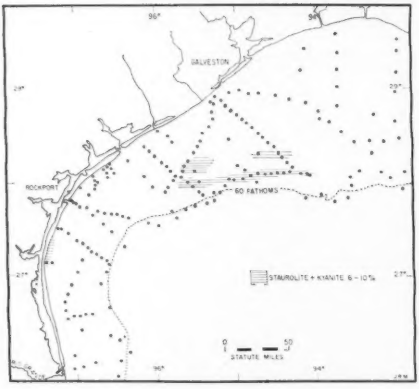

FIG. 11.—Areal distribution of kyanite+staurolite percentages in the Western Gulf province; 89 percent of samples conform to pattern.

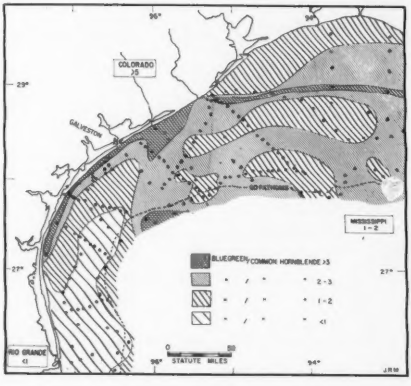

FIG. 12.—Areal distribution of hornblende ratio (ratio blue-green/common+brown hornblende) in the Western Gulf province; 92 percent of samples conform to pattern.

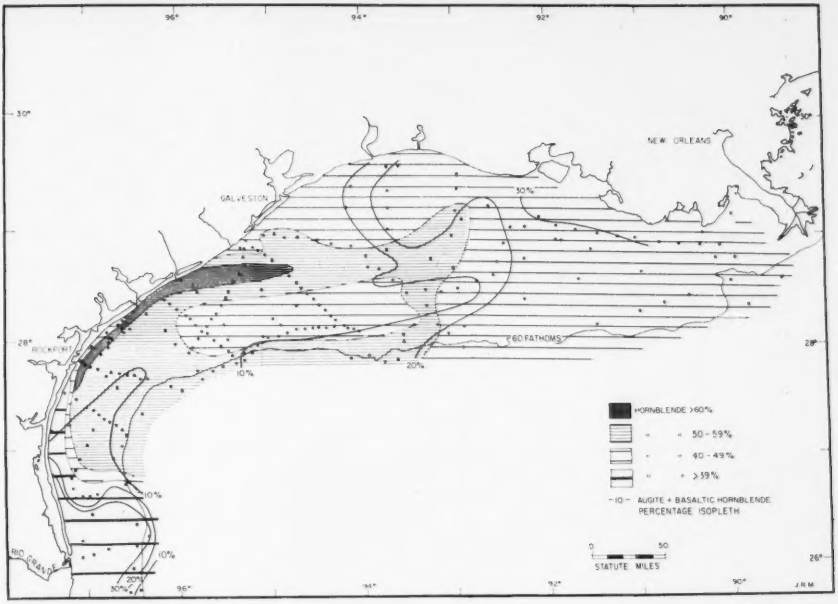

FIG. 13.—Areal distribution of hornblende (not including basaltic) and pyroxene+basaltic hornblende percentages in the Western Gulf province; 86 percent of hornblende percentages and 98 percent of pyroxene+basaltic hornblende percentages conform to pattern.

province and indicates in order of quantitative importance the sources of the sediments in each zone. The conclusions that can be drawn from figures 10–14 can be summarized as follows:

1.—The values of all parameters are lower in the Western Gulf province than in the sediments of the rivers that have contributed the material, thus indicating that all Western Gulf sediments are mixtures in varying proportions of two or more components. Only in a narrow belt close to the shore between the mouth of the Colorado River and Corpus Christi and in the area north of the modern Rio Grande province relatively pure associations derived from the Colorado and from weathered Rio Grande deposits respectively can be observed.

2.—North of the modern Rio Grande province and west of the Mississippi province the sediments appear to have been derived predominantly from older Mississippi and Rio Grande deposits which have lost their pyroxenes by weathering.

3.—In the central part of the Western Gulf province the assemblages occur in belts parallel to the shelf edge or the coast line. Zones of Colorado material or mixed detritus derived from the Colorado, Brazos, and Trinity Rivers alternate with sediments in which the Mississippi element predominates. This zonation is clearly illustrated by figure 15. On the shelf south of the mouth of the Colorado River the zonation is interrupted by a north-south trending belt of almost pure Colorado material extending across the entire width of the shelf and beyond the shelf edge.

4.—Elements derived from Texas rivers exclusive of the Rio Grande and Colorado, or reworked from the Pleistocene of the Texas coast, are concentrated in three areas (figs. 10 and 11). Two of them are situated parallel and relatively close to the shelf edge in the area south of Galveston and the Colorado River. The high zircon percen-

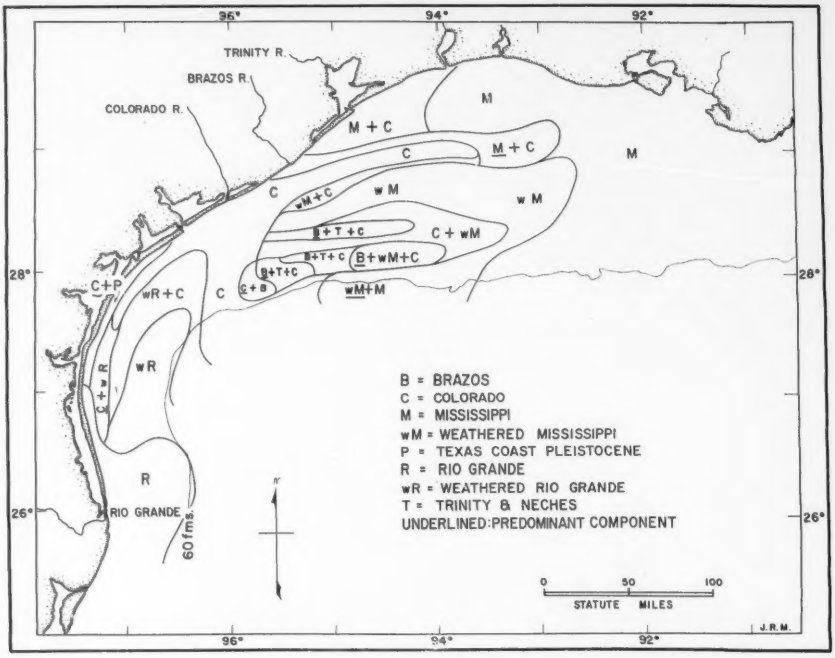

FIG. 14.—Sketch map of sources and mixing of sediments in the Western Gulf province. Boundaries are approximate and based upon figures 10–13. Components for each area are listed in order of quantitative importance.

tages and low hornblende/tourmaline ratios demonstrate considerable influence of Texas river material, probably mainly the Brazos. The influence of the Trinity and Neches Rivers is noticeable in relatively high kyanite and staurolite percentages. The third zone of high zircon percentages and low hornblende/tourmaline ratios occurs along the present coast from Baffin Bay to a point just west of the mouth of the Colorado River. Given the insignificance of the rivers in this area it appears probable that these sediments have been derived largely from outcropping Pleistocene materials. The question of their origin will be discussed further in the next section.

5.—There is no Brazos distributive province off the mouth of this river of a size observable with the available sampling grid. Nienaber (1958) has demonstrated that pure Brazos sediments, relatively uncontaminated with Mississippi and Colorado material, are restricted to a small area in the immediate vicinity of the river mouth. On the other hand, a zone of almost pure Colorado material extends westward from the mouth of this river in an elongated belt parallel to the coast.

The reasons for the mineral distribution pattern shown in figure 14 are complex. Present day deposition of heavy minerals appears to be a factor of importance only in the zone immediately along the coast and off the Rio Grande and Mississippi Rivers. Everywhere else the pattern is controlled by the interplay of three main factors: the post-Pleistocene rise in sea level from a low stand approximately at the edge of the shelf; shifts of current and longshore drift patterns in the Gulf caused by climatic changes during early Holocene time; and the pre-existing drainage patterns of the rivers on the ex-

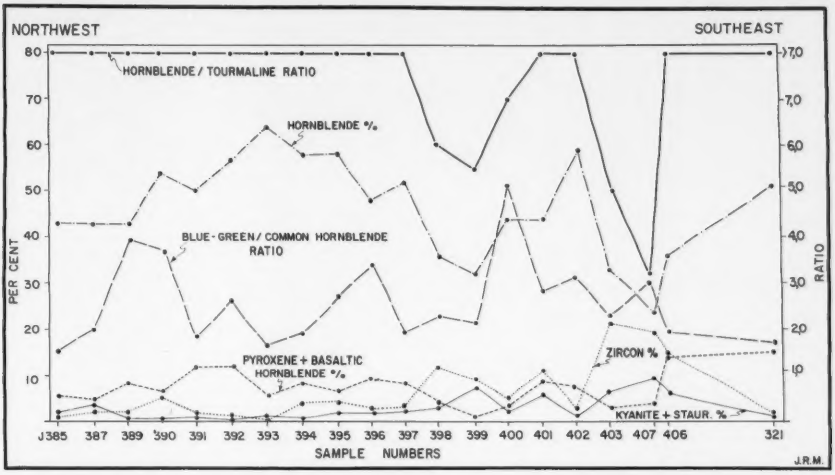

FIG. 15.—Graphic cross section of heavy mineral variation across the continental shelf east of Galveston. For location of section see figure 1. For significance of parameters see table 10.

posed continental shelf. The latter may have been controlled at least in part by tectonic movements.

Curray (in press) has analyzed in detail the Holocene history of the continental shelf on the basis of bathymetrical observations, grain size analyses, and the heavy mineral data of the present paper. A brief summary of this history in terms of sediment sources follows here (fig. 16).

During the period of lowest sea level stand, sediments from all rivers were carried directly downslope across the exposed shelf to deep water. A marine current and drift pattern similar to the one prevailing today is suggested by the northward extension of Rio Grande material and the westward trending tongue of Mississippi sediment. Material from the Colorado, Brazos and Trinity Rivers was deposited in a neutral zone between the areas of eastward and westward drift (fig. 16, stage I).

During the time of early sea level rise this distribution pattern changed appreciably as a consequence of a major reduction in westward drift of the Mississippi material. First, the sediments derived from the Brazos and Trinity Rivers, and a little later also those from the Colorado, were carried eastward over an appreciable distance in channels that are still partially exposed on the shelf (fig. 16, stages II and III).

In the next interval (stage IV) the predominant westward drift of Mississippi sediment and the deposition of Colorado, Brazos and Trinity material in a neutral zone were restored. Rio Grande sands continued to be moved northward as in all previous stages. Mississippi and Rio Grande sediments deposited in the Western Gulf province in this and earlier stages have apparently been derived from weathered deposits or were weathered in place after deposition because they are poor in pyroxene. Younger sediments from both sources contain normal amounts of this mineral.

When sea level reached the 10 fathoms depth contour (stage V), another major shift in the directions of transport occurred. Again the westward drift was reduced and the Colorado and Brazos sediments were carried far east into the area south of the Calcasieu River. How far east the Colorado deposits extended is not known, because in the area east of the Calcasieu River all older deposits are obscured by a blanket of modern Mississippi sediment. At this time Rio Grande sediment was deposited as far north

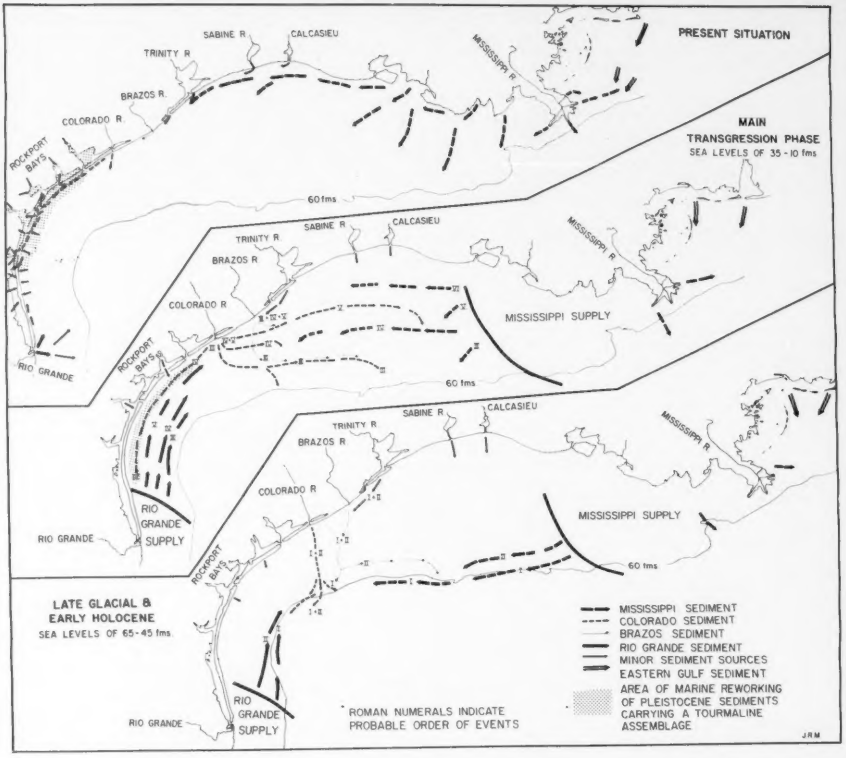

FIG. 16.—Sediment sources and directions of transportation on the continental shelf of the northern Gulf of Mexico during the Holocene. Roman numerals indicate sequence of events but do not represent a correlation with any stratigraphic system for the Holocene period.

as the area of Rockport.

Major climate changes are assumed to be responsible for the marked change in transport direction in stages II, III, and V.

When sea level rose above the 10 fathom contour the predominant westward drift was restored again (fig. 16, stage VI) as reflected by another shift of the sediments derived from the Mississippi. Virtually the same transportation pattern characterizes present day deposition although, as will be shown in the next section, westward drift is probably not as prominent. At the present time deposition of sand-size material seems to be restricted to the shelf off the Mississippi and the Rio Grande and to a narrow nearshore zone between those rivers. Except for some westward drift of Colorado and Brazos material as far as the area of the Rockport bays, sediments in this nearshore zone appear to have local sources.

Thus, in the Western Gulf province the mineral distribution pattern of the basal part of a transgression has been preserved. It is of interest to note that this pattern is determined almost entirely by nearshore distribution processes operating during various stages of the transgression and has not been significantly modified by the deep water environments following them. Only in the

areas immediately off very large rivers do sand-size sediments spread at the present time across the entire width of the shelf.

HEAVY MINERAL DISTRIBUTION, COASTAL ZONE BETWEEN BRAZOS AND RIO GRANDE RIVERS

The mineralogy of the coastal and nearshore sediments of the northwestern Gulf of Mexico has not been examined as systematically as that of the continental shelf deposits. Detailed sample collections are available for Laguna Madre, Aransas, San Antonio, and Mesquite Bays. A few samples have been investigated from Nueces, Copano, and Galveston Bays, and a satisfactory collection is available from Padre, St. Joseph, and Matagorda Islands. No information at all exists for the zone between Galveston Bay and the Mississippi delta.

Five heavy mineral associations can be expected in the sediments of the Texas coastal region: the pyroxene-hornblende association of the Rio Grande, the hornblende association of the Colorado, the tourmaline association of the Brazos and the rivers between the Colorado and Rio Grande and of the Pleistocene, the tourmaline-kyanite-staurolite association of the Trinity, Neches and Sabine Rivers, and the hornblende-pyroxene association of the Mississippi. A large part of the deposits probably consists of mixtures of two or more components.

The heavy mineral variation along the beaches of Texas has been described by Bullard (1942). His principal conclusions follow (fig. 17). The sediments of the beaches and barrier islands have been transported west and southwestward by longshore drift except in the area just north of the Rio Grande where indications of northward movement have been found. The rivers are considered the only important source of sand and the contribution from the erosion of the headlands is supposed to be negligible. Nearly all sediments west of the mouth of the Colorado are supplied by this river although Nueces material is considered an important diluting element in the area of Corpus Christi Bay and the northern part of Laguna Madre. The predominantly stable assemblages occurring east of the Colorado are considered a direct contribution from the Brazos River diluted with material from the Mississippi supplied by longshore drift.

The results of the present study suggest that these conclusions of Bullard require considerable modification.

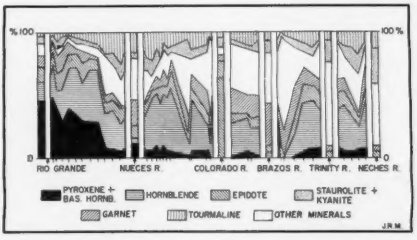

FIG. 17.—Heavy mineral variation diagram for the Texas Gulf Coast beaches between the Rio Grande and the Neches Rivers. Data converted from Bullard (1942). For sample locations see Bullard (1942, fig. 1). Marks on base line are sampling points.

Laguna Madre

In the southern part of Laguna Madre (fig. 18, table 11) the sediments are characterized by a pyroxene-hornblende associa-

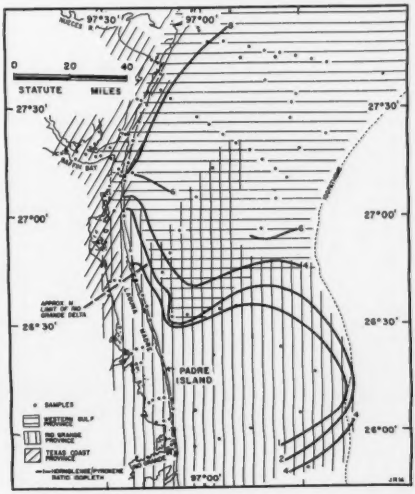

FIG. 18.—Regional distribution of heavy mineral assemblanges in Laguna Madre and on the adjacent continental shelf.

TABLE 11.—*Representative analyses of samples from Laguna Madre, Corpus Christi Bay, and the adjacent continental shelf, in percent of total non-opaque fraction*[1]

Sample number	Hornblende[2]	Tourmaline	Epidote	Zircon	Garnet	Staurolite	Kyanite	Titanite	Basaltic Hbl.	Augite	Hypersthene	Others
Rio Grande Province												
LM 20	22	7	23	7	7	4	4	3	5	18		
65	15	4	19	9	21	2	4	6	5	13	2	1
69	18	5	19	7	13		1	5	1	28	2	1
75	21	3	19	1	10	4	3	5	9	24		1
76	28	8	19	6	4	5	1	4	10	15		
J 479	22	1	14	17	17	1	3	1	7	10	4	3
718	33	2	8	7	7		1	8	3	31		
723	17		4	10	14	3	1	4	3	44		
729	23	1	12	7	11			4	4	34	4	
732	38		4	4	2	1	1	3	5	41	1	
737	29	2	5	9	6	1	1	3	7	37		
739	28	1	19	8	6			1	2	28	5	2
Texas Coast Province (Laguna Madre)												
LM 2	7	32	6	34	4	8	4					5
3	8	22	8	36	4	7	8					7
6	3	39	11	7	8	21	8		1			2
7	1	32	21	18	5	14	5					4
8	5	26	13	29	4	10	5	1				6
11	3	28	18	18	9	7	6	2		1		8
Texas Coast Province (Nueces Bay)												
N 1	5	29	14	16	11	9	6	3		2		5
2	4	32	10	18	9	6	3	3		3		12
3	12	23	10	18	7	8	6	3		1		12
4	1	30	8	14	12	16	3	3	1	3		9
Mixed assemblages												
LM 57	9	8	17	21	25	4	2	6	4			4
78	16	20	12	20	10	4	3	4		2		9
79	6	21	4	36	10	9		4	1	2		7
88	15	19	8	36	4	3		3		2		10
120	36	27	15	4	6	7		1				4
123	32	23	11	11	4	6	2	1	3	3		4
203	2	13	29	15	15	11	3	2		8		2
205	11	9	27	18	17	3	1	3	5	4		2

[1] For sample location see fig. 1, and Rusnak (in press) fig. 27.
[2] Not including basaltic hornblende.

tion and form part of the Rio Grande province. The northern limit of this province in Laguna Madre and on the adjacent shelf coincides almost precisely with the morphological limit of the pre-modern Rio Grande delta (Rusnak, in press). At the present time hardly any sediment from the Rio Grande is brought directly into the lagoon. Consequently, the pyroxene-hornblende assemblages of the southern part of the lagoon must have been derived largely from reworked older deltaic deposits. The transitional zone between the Rio Grande province and the Texas Coast province following

TABLE 12.—*Heavy mineral assemblages from the late Pleistocene (Beaumont Clay) of the Texas Coast, in percent of total non-opaque fraction*[1]

Sample number	Hornblende[2]	Tourmaline	Epidote	Zircon	Garnet	Staurolite	Kyanite	Titanite	Basaltic Hbl.	Augite	Hypersthene	Others
Laguna Madre												
LM 91C	3	29	9	26	6	3	6	4		1		3
97B	13	26	11	18	7	9	7	2		1		6
110A	11	19	12	32	6	6	6	1	1	1		5
189A	1	20	6	48	6	10	4					5
198C		30	9	29	10	7	8	3	1	1		2
213A	25	6	19	6	9	3	6	5	6	14		1
Rockport area												
RA 1	8	18	4	49	4	6	2					9
3	48	27	6	9	4	2	1					3
4	32	19	13	18	3	7	2	1				5
5	23	45	5	14	6		1	2				4
6	7	32	7	20	10	9	4	3				8
Houston (After Rogers and Powell, 1958; Table 1, average values)												
		34		53	3	1	2					7

[1] For sample locations see Fig. 21 and Rusnak (in press), fig. 13.
[2] Not including basaltic hornblende.

to the north is relatively narrow. Apparently, little sediment is at present being transported along the axis of the lagoon.

In the northern part of Laguna Madre, in Corpus Christi Bay, and Baffin Bay the sediments on the land side are characterized by a tourmaline association with varying but generally small amounts of hornblende. This tourmaline association is similar to the assemblages of most smaller Texas rivers (table 3) and also to the heavy mineral suites of the Pleistocene of Laguna Madre and the Rockport Bays (table 12). No rivers of importance drain into Baffin Bay, and the northern Laguna and the distribution of the tourmaline association along the inland shores suggests that reworked Pleistocene may be the principal source of the material. Modern Nueces sands contain small but significant percentages of volcanic minerals (table 3). These minerals are found only in the upper part of Nueces Bay close to the river mouth. Apparently, this river does not contribute much sediment to the outer part of the bay or to the northern Laguna area.[3]

In its purest form the tourmaline association is found on the land side of the northern Laguna Madre. Seaward across the lagoon

[3] Two samples of Pleistocene beach rock from the western shore of Laguna Madre just north of the entrance to Baffin Bay have been investigated to test the possibility that part or all of the Pleistocene material in this area may represent old Rio Grande delta sediment that has lost its unstable minerals by weathering. In the beach rock such unstable components might have been protected by the carbonate cement. One of the samples contains a normal Texas coast assemblage (hornblende 21 percent, tourmaline 15 percent, zircon 35 percent, kyanite plus staurolite 9 percent, hornblende ratio 1.9) entirely within the range of modern assemblages from this area. The other is also similar to modern samples except for the presence of 2 percent of very badly weathered but recognizable pyroxene and approximately 30 percent of deeply altered grains now consisting of fine micaceous aggregates. These grains might represent completely altered pyroxenes and would thus indicate that the source of this material is a Pleistocene Rio Grande deposit. However, the hornblendes appear to be almost completely fresh and the hornblende ra-

and the barrier island the percentage of hornblende increases rapidly. This hornblende is to a large extent of the blue-green variety typical for Colorado material, suggesting that the ultimate source of at least part of this material is the Colorado River. This material may either have been transported southwestward by longshore drift as suggested by Bullard (1942) or derived from some older deposit of Colorado material in the vicinity. With the aid of the hornblende/tourmaline ratio it can be estimated that inside Laguna Madre more than 70 percent of the heavy minerals belong to the tourmaline province, and hence were derived locally. The small variable amounts of hornblende present in the Pleistocene sediments make it impossible to obtain a more accurate estimate. The proportion of locally derived material in the total sediment is much higher because Colorado sediments contain at least three times as many heavy minerals per unit sand as the Texas Coast province. Consequently, in a mixture the contribution of Colorado material is heavily over-emphasized.

In the discussion of the Western Gulf province use has been made of several parameters characterizing the various sediment sources. The same parameters can be used to unravel the mixtures of sediments in the coastal zone. The parameters used are the hornblende/pyroxene ratio, measuring the relative proportion of Colorado versus Rio Grande or Mississippi material; the hornblende/tourmaline ratio indicating the proportion of Pleistocene or Texas river-sediment; and the hornblende ratio, which gives the proportion of Colorado sediment. For characteristic values in the pure source materials see table 10.

The northern limit of the Rio Grande province is indicated by a sharp increase in the hornblende/pyroxene ratio (fig. 19). Inside the lagoon this increase coincides with the morphological limit of the pre-modern Rio Grande delta but on Padre Island the increase is shifted a little to the north in-

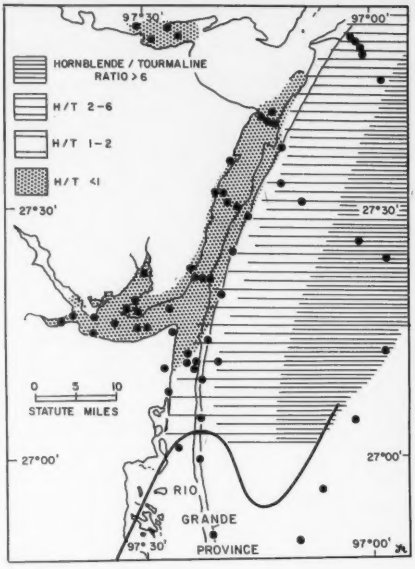

FIG. 19.—Variation of mineral parameters indicating sediment sources in bay and barrier islands deposits between the Rio Grande and Galveston Bay. Black dots indicate barrier island samples; open circles indicate bay samples. curves indicate trends only.

dicating a limited amount of northward longshore drift. North of the Rio Grande delta blue-green hornblende is an important component of the hornblende fraction, but the hornblende ratio remains considerably below the value of 5.0 typical for pure Colorado material (fig. 19). The variety which lowers the value must have been derived from the Rio Grande. Direct northward transportation along the shore appears to be excluded because of the low percentage of associated pyroxene which renders it unlikely that modern Rio Grande material is contributed to this area. The source of the hornblende assemblage which dilutes the tourmaline association in the northern Laguna Madre must be a mixed deposit of Colorado and weathered Rio Grande material. The nearest source of such detritus is found on the continental shelf where according to figures 12 and 13 sediments with the correct pyroxene content and horn-

tio of 21 places the sample entirely within the range of modern northern Laguna material, as contrasted with a ratio of 1.0 or less for Rio Grande material. Although this appears to refute the Rio Grande as a source for this sample, the question deserves further study.

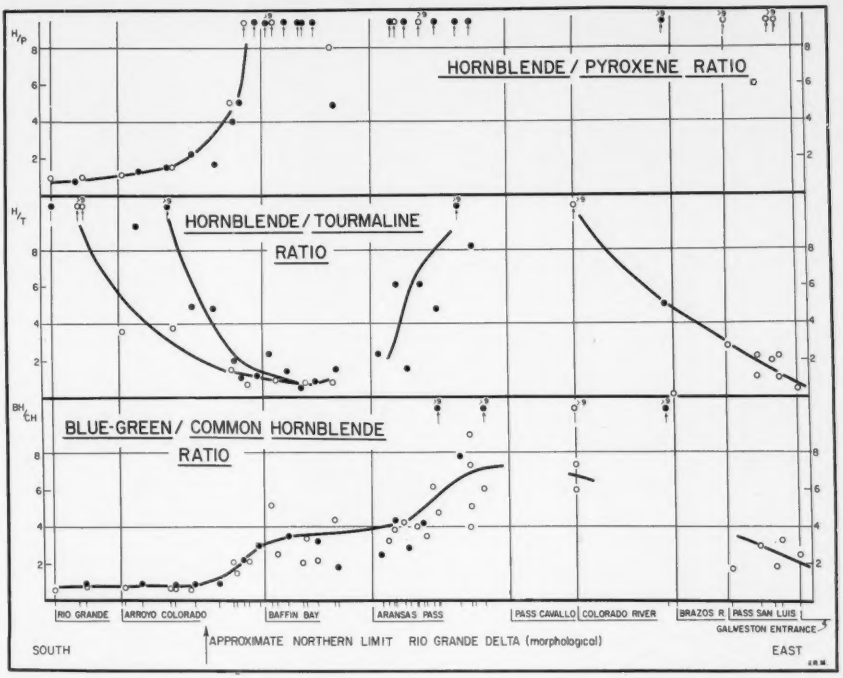

FIG. 20.—Areal distribution of hornblende/tourmaline ratios indicating relative contributions of Colorado and Pleistocene material in the northern Laguna Madre and on the adjacent continental shelf.

blende ratio are present over an extensive area.[4]

Neither the amount of hornblende in the assemblages nor the composition of the hornblende fraction change significantly between the northern limit of the Rio Grande province and Aransas Pass (fig. 19). If, as Bullard assumes, longshore drift in a southwestward direction is the principal mechanism of hornblende supply, a southwestward decrease of this mineral should be expected. This apparently is not the case. The importance of the hornblende admixture increases away from the shore rather than parallel or oblique to it (fig. 20), suggesting, as does the constant presence of Rio Grande components over the entire area between the Rio Grande delta and Aransas Pass, that the shelf off this part of the coast is the source of the hornblende suite.

Contrary to Bullard, it must be concluded that the sediments of the bays and barrier islands between the Rio Grande and Aransas Pass were not brought there by longshore drift with local dilution of river sands, but were locally reworked from preexisting deposits on land and on the continental shelf. Net transportation, instead of parallel to the coast, was predominantly perpendicular to it and probably associated with the Holocene transgression. The prin-

[4] An estimate of the amount of Rio Grande sediment present in the assemblage diluting the tourmaline association in Laguna Madre can be obtained with the aid of the characteristic hornblende ratios of the pure Colorado and Rio Grande associations. The hornblende association appears to consist of approximately equal portions of Colorado and Rio Grande sediment, assuming that the heavy mineral contents of both types of material are roughly the same.

TABLE 13.—*Representative heavy mineral analyses of samples from Galveston, Aransas, Mesquite and San Antonio bays, Texas, in percent of total non-opaque fraction*[1]

Sample number	Hornblende[2]	Tourmaline	Epidote	Zircon	Garnet	Staurolite	Kyanite	Titanite	Basaltic Hbl.	Augite	Hypersthene	Others
Galveston Bay												
GB 1	41	15	21	8	10	1	1		1	1		1
4	3	47	2	17	17	10	2					2
5	3	42	8	20	12	3	3					9
16	31	39	13	2	5	4	3	1				2
Aransas Bay												
A 22	76	10	8		2	2		1				1
31	78	8	8		3	3						
40	44	29	15	3	5	2	1					1
60	38	36	11	1	1	5	5	1				2
77	9	58	13	5	1	10	2					2
Mesquite Bay												
M 7	60	13	17	2	2		2	1	1	2		
9	19	56	9	1	2	8	3		1			1
15	61	13	17	3	3	2		1				
22	59	20	15	2	2	1		1				
25	4	52	20	3	5	6	5	1	2	1		1
40	28	51	11	2		3	3					2
44	64	11	13	2	3	3	1	2				1
San Antonio Bay												
S 16	68	16	8	1	1	2	1	1				2
34	38	23	12	12	5	2	2	1	2			3
92	8	38	7	18	7	5	7	2				8
95	22	30	10	15	7	7	4	1	1			3
110	42	27	7	8	7	2	4	2				1
111	64	13	8	10	5							
113	52	30	8	1	3		3	2				1
141	7	37	18	22	4	5	5	1				1
168	44	37	5	1	2	4	2	2		1		2
188	11	44	15	13	7	3	1	2			2	2
191	50	17	11	9	4	3	1	1	1		1	2
270	31	51	3	4	3	4	2					2
340	55	21	8	4	7	2	1	1			1	

[1] For sample locations see fig. 1, and Poole (1958), figs. 2 and 3.
[2] Not including basaltic hornblende.

cipal contributors of sediment are old Rio Grande deposits for the southern part of Laguna Madre, and Pleistocene deposits for the northern Laguna, Corpus Christi Bay, northern Padre Island, and Mustang Island. Mixed Colorado and weathered Rio Grande deposits of early Holocene age on the shelf contributed a mineralogically conspicuous but quantitatively minor amount of material.

Rockport Area, Texas

East of Aransas Pass both the total amount of hornblende and the hornblende ratio increase rapidly in the direction of the Colorado River (fig. 19). Midway between

TABLE 14.—*Representative analyses of samples from Padre, St. Joseph, and Matagorda islands, in percent of total non-opaque fraction*[1]

Sample number	Hornblende[2]	Tourmaline	Epidote	Zircon	Garnet	Staurolite	Kyanite	Titanite	Basaltic Hbl.	Augite	Hypersthene	Others
Padre Island												
P 54	24	4	17	8	13	1		3	3	25		2
48	40	4	11	2	6	1	2	1	12	20		1
42	28	5	17	1	9	4	4	1	10	20		1
33	29	6	20	6	10	2		6	9	9		3
30	36	26	6	2	8	2	4	4	5	5		2
27	32	31	8	6	5	4	3	1	4	2		4
21	51	21	13		3	6	3	2		1		
18	41	29	7	3	3	5	4	3		2		3
St. Joseph Island												
SJ 5	51	22	14	3	7	2			1			
20	61	10	6	6	6	1	3	3	1			3
36	61	12	18	3	1	1		1	1			
51	43	16	12	16	4	4	2				2	1
53	41	28	12	1	6	3	2	2	2		3	
Matagorda Island												
MT 11	53	16	13	5	2	5	2				1	2
17	55	9	16	6	6	5	1	2				1
26	62	13	9	3	4		2	1	1	2	1	2
33	72	4	5	5	5	3	2	2	2			
42	65	8	8	8	6	2	1	1	1			
45	59	10	11	5	5	6		2				2
53	57	16	14	2	8	2		1				
62	34	13	11	30	6	2	2					2
71	56	14	12	1	6	5	2	2				2
Sargent Beach (just west of Brazos mouth)												
	28	6	4	38	10	2	2	2	3			5

[1] For sample locations see figs. 1 and 21.
[2] Not including basaltic hornblende.

Aransas Pass and Pass Cavallo (an old Colorado mouth) the coastal sediments are almost entirely derived from the Colorado River. On St. Joseph and Matagorda Islands and in the nearshore zone the eastward gradient caused by gradual replacement of the tourmaline association by deposits rich in Colorado hornblende is very pronounced. The influence of strong westward longshore drift in this area is well substantiated.

Quite extensive heavy mineral information is available for the bays of the Rockport region (Poole, 1958). Two associations, a hornblende and a tourmaline association (tables 13 and 14), and their mixtures in varying proportions can be distinguished. The main components, hornblende and tourmaline, are almost mutually exclusive in the pure assemblages and together make up approximately 75 percent of all suites. Therefore, a map of the hornblende/tourmaline ratios accurately reflects the distribution and mixing of heavy mineral suites (figs. 21, 22).

In its purest form the tourmaline assem-

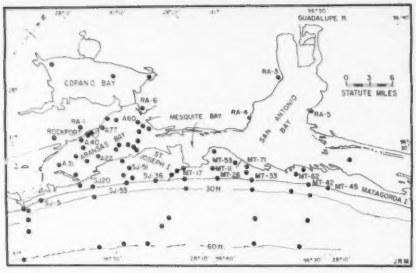

FIG. 21.—Place names in Rockport bays and sample location for Aransas Bay, the barrier islands and the adjacent continental shelf. For sample locations in Mesquite and San Antonio bays see Poole (1958, figs. 2, 3).

blage is found in a broad belt along all landward shores of the bays and also off the deltas of the rivers emptying in the bays. Both the rivers and the local Pleistocene deposits (tables 3 and 12) carry this association. The regional distribution suggests that erosion of the Pleistocene shore is probably the most important source of this suite.

Toward the center of the bays and in the direction of the barrier islands the hornblende/tourmaline ratio increases gradually as a result of dilution with the hornblende association. The hornblende ratio in these deposits shows that the hornblende has been derived from the Colorado River (fig. 23).

In detail the mineral distribution patterns of figure 22 are complex and of areally unequal value because of varying sample density. The main trends, however, are clearly established. There is a marked westward decrease of the Colorado component which is also reflected in the nearshore deposits. Farther offshore on the shelf the zone of modern Colorado sediment merges with an older zone of sediments from the same source (see fig. 14). Apparently, the influence of the Colorado River does not extend as far west in the nearshore zone as during the last phase of the Holocene transgression.

A sharply defined tongue of Colorado sediment extends from the near-shore Gulf across the barrier island into San Antonio Bay approximately at the site of a former major inlet. A similar tongue can be observed in Aransas Bay. Apparently, the supply of Colorado sediment to the bays is caused by a sharply localized stream of sediment rather than to washover by storm waves over a broad front, or any other mechanism involving temporary deposition in the barrier island. No Colorado material in appreciable quantity has entered San Antonio Bay from the river by way of Espiritu Santo Bay. Apparently transportation of sand size material between bays is restricted.

Recently, Parker (1959) has postulated a Holocene estuarine phase for San Antonio Bay during which there was an open connection between the bay and the Gulf, possibly similar to the situation at Calcasieu Lake. This would explain the localization of the Colorado sediment tongue as a remnant of an open phase of the bay. Unfortunately, the available subsurface heavy mineral information, although suggestive of a correlation between the first appearance

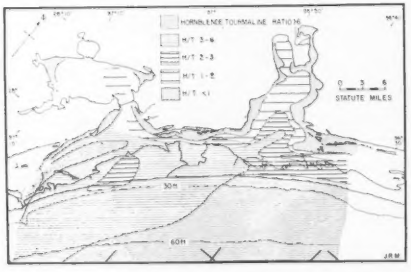

FIG. 22.—Areal distribution of hornblende/tourmaline ratios in the bays of the Rockport area; 95 percent of samples conform to pattern. For bay names see fig. 21.

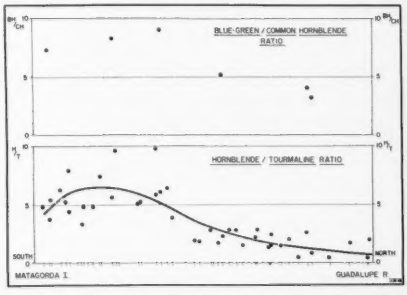

FIG. 23.—Variation of hornblende/tourmaline ratio and hornblende ratio along axis of San Antonio Bay.

of the Colorado tongue and the estuarine phase, does not permit a definite conclusion.

The actual proportion of Colorado sediment in the deposits of the bays cannot be inferred from the heavy mineral assemblage alone. The heavy mineral content per unit sediment of the tourmaline assemblage is not known, but by analogy with other areas in the Texas Coast province it may be assumed that it is at least a few times lower than that of the Colorado sediment. Consequently, the importance of the Colorado contribution to the bay sediments is exaggerated by the heavy minerals.

East Texas Coast

East of Pass Cavallo the information becomes very scarce. Bullard's data (fig. 17) show that as far east as the mouth of the Brazos River relatively high percentages of hornblende persist. This hornblende probably is not derived directly from the Colorado but reworked from older deposits in this area where the mouth of the river has repeatedly migrated. More than 60 percent of the heavy mineral assemblage in this area, however, consists of stable minerals probably supplied by the Brazos and brought by westward longshore drift. The submarine Brazos delta is of very restricted size and contains a mixture of Colorado, Mississippi, and Brazos components (Nienaber, 1958). Only along the shore pure Brazos material is found.

East of the mouth of the Brazos another increase of the hornblende percentage occurs together with a slight increase in the amount of pyroxene. Bullard (1942) has explained this by assuming that Mississippi sediment was carried westward by longshore drift. However, the hornblende/pyroxene and hornblende variety ratios in this area both exceed by an appreciable amount the typical values of the Mississippi association. At least part of this material must, therefore, have been derived from the Colorado. To explain this mixture it is not necessary to assume periodic or temporary replacement of the westward drift by an eastward movement, because not far offshore on the shelf a zone of relatively pure Colorado sediment was deposited during the later part of the Holocene transgression. This zone may have supplied the Colorado element in the coastal deposits, whereas westward drift accounts for the supply of Mississippi material.[5] The mixed assemblages of hornblende and tourmaline are restricted to the barrier islands, beaches, and outer bays. Inside Galveston Bay (fig. 19, table 13) the sediments are characterized by relatively pure assemblages of the tourmaline association derived either directly from the Trinity and Neches Rivers or from older deltaic deposits.

In conclusion it can be said that the sediments of the Texas Coast province are largely of local origin and formed by reworking and dispersal of older deposits present in the coastal zone itself or in the immediate vicinity on the continental shelf. Net long distance transport of Colorado material in the nearshore zone plays an important role in the nearshore area east of Aransas Pass, and its products are being accumulated mainly in the beaches, barrier islands, and outer parts of the bays. Although the Colorado element is conspicuous and quantitatively important in the heavy mineral assemblages, it can be assumed that its importance in the total sand fraction of the sediments in the bays is considerably less.

CONCLUSIONS

The present paper deals with the sources of the sands and the sand fraction of the clays in the northern Gulf of Mexico; its conclusions apply only to this size fraction.

A total of seven sources contributes sediment to the northern Gulf of Mexico: the rivers draining the sedimentary mantle of the southern Appalachians, the Mississippi, the rivers of western Louisiana and east

[5] Price (1958) has suggested the possibility that parts of the heavy mineral assemblages in this area may have been derived from old Brazos deposits formed before an assumed capture of the upper course of this river by the Pecos. They would therefore at least partly consist of Rocky Mountain material. Sidwell (1941) has shown that the modern Pecos carries an epidote-tourmaline assemblage. It is possible that some of the tourmaline and epidote in samples from the Galveston area may ultimately have been derived from the headwaters of the modern Pecos, but in view of the evidence for supply of a Mississippi association (based on pyroxene and hornblende, both minerals absent in the Pecos) this does not seem to be a necessary assumption.

Texas, the Colorado, the Brazos and the rivers between the Colorado and the Rio Grande, the Rio Grande, and the Pleistocene deposits of the coast of Texas. Only four of these, the Mississippi, Brazos, Colorado and Rio Grande, are quantitatively of great importance.

In addition to the incompletely studied Eastern Gulf province, four major distributive provinces can be recognized. Two of these, the Mississippi and Rio Grande provinces, have vast and distant source areas. Active deposition of material from the Mississippi extends at the present time from the subaerial delta to the edge of the shelf and beyond. The mineral composition of this province is homogeneous over a very large area. The dispersion of Rio Grande material over the shelf is at the present time probably much more restricted. Beyond the 10 fathom depth contour much of the material in this area may not represent present day deposition.

The Western Gulf province is of complex origin and at present largely inactive as far as the deposition of sand-sized material is concerned. Its formation is associated with the early Holocene transgression and the sediment distribution is primarily caused by nearshore processes. The sediments are derived from virtually all rivers in or near the area, including the Mississippi, partly directly and partly by reworking of older deposits that suffered considerable weathering during the last glacial period. The sediment distribution within this province has been modified very little by the present deep-water environment and essentially is a relict of successive nearshore environments associated with the transgression.

The Texas Coast province is restricted to a narrow belt in the coastal area of Texas and probably also of western Louisiana. Its sediments have been derived largely from local marine erosion and reworking of Pleistocene and early Holocene deposits, partly in connection with the Holocene transgression. In the eastern part of this province, between Galveston Bay and the Rockport bays, longshore drift plays a role in the transportation of sediments, but the quantitative importance of long distance sediment supply is not great.

Radial sediment distribution patterns prevail off the large Rio Grande and Mississippi Rivers. The mineral assemblages around such powerful sediment sources are homogeneous over large areas. Between these loci of active sediment supply, sediment dispersion patterns are complex, and predominantly parallel to the margins of the basin. Sediment supply is less vigorous, and sedimentation is transgressive rather than regressive. Transportation in this zone is caused by nearshore processes, and marine redistributing agents prevail over the forces that tend to build deltas from the shore.

Geologically, the sediment thickness and time interval involved in the present study are small. In an ancient basin the sequence studied here would probably be accepted as an approximately simultaneous series of events. In the light of the results of this study the great difficulties encountered in long range heavy mineral correlation are easily understood.

ACKNOWLEDGMENTS

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FIGS. 1–12, MARCH, 1960

RELATIONSHIP OF THE INSULAR SHELF SEDIMENTS TO THE SEDIMENTARY ENVIRONMENTS AND GEOLOGY OF ANACAPA ISLAND, CALIFORNIA[1]

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ABSTRACT

Anacapa Island comprises three small islands which lie approximately 12 miles off the southern California coast along longitude 34°00′30″ N. They rise above a submerged westward-trending anticlinal ridge which extends seaward from the western terminus of the Santa Monica Mountains. The island is barely 5 miles long and includes an area of about 1 square mile.

Gentle north-dipping Middle Miocene andesitic lavas and pyroclastic rocks of the Conejo Volcanics form the bulk of the island. Two strata of the San Onofre Breccia, each a maximum of 35–40 ft thick, occur as interbeds near the base of the exposed volcanic rocks which are about 1700 ft thick. Fossiliferous marine sands 2–3 ft thick of Early Pleistocene(?) age and alluvial sands 25–30 ft thick of Late Pleistocene age overlie a summit terrace atop the central of the three islands. The Conejo Volcanics crop out over much of the insular shelf east, west, and south of the island. Siliceous and calcareous shale and gray mudstone of the Monterey Formation of Middle Miocene age crop out about one mile west of the island beneath Anacapa Passage.

Two contrasting sedimentary environments, the geology of the island and the surrounding shelf, and the submarine topography control the character and distribution of the unconsolidated shelf sediments. East and west of the island relatively strong currents prevent the accumulation of detrital clasts and the sediment blanket is thin and mainly composed of calcareous animal and plant fragments. In areas of submarine outcrops relict and residual detrital sand and gravel have formed. North and south of the island detrital sedimentation is dominant; the deposits decrease in grain size with distance from shore and are composed mainly of sand size rock and mineral fragments derived from the Conejo Volcanics and to a lesser extent from the San Onofre Breccia. Detrital clasts finer than about 0.125 mm are able to by-pass the rather steeply sloping (1°30′) and narrow southern shelf. A much wider and more gently sloping northern shelf largely prevents by-passing of the finer sand clasts north of the island. Sedimentation has been fastest on the northern shelf during Recent time, and the sediment blanket is here thickest and most uniform.

INTRODUCTION

Anacapa Island is one of the smallest of eight islands which rise above a foundered segment of the continent off southern California known as the Continental Borderland (Shepard and Emery, 1941). Until this study Anacapa Island had not been mapped geologically and little was known of its shelf sediments. Anacapa Island is unique among the offshore islands because persistent wave-attack during Recent time has destroyed so much of the island that complete wave-leveling is foreseeable in the near geologic future. Moreover, the small size of this island in comparison with a much larger surrounding shelf affords excellent opportunities for the study of Recent insular shelf sediments in relationship to the sedimentary environments and geology of the island.

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Anacapa Island lies nearly 12 miles off the southern California coast along latitude 34°00′30″N. and is the easternmost member of the four Santa Barbara channel islands (fig. 1). The island trends east-west and is approximately bounded by longitudes 119°21′ W. and 119°27′ W. The only permanent inhabitants operate a United States Coast Guard lighthouse at the eastern tip of the island. Anacapa derived its name through repeated misspellings of the Chumash Indian word, Enneapah, which is thought to mean deception.

Previous Work

After the sinking of the steamship Winfield Scott which struck Anacapa in the winter of 1853, interest was aroused in the island as a possible lighthouse site. Lieut. T. W. Stevens' report on this site in 1854 contains the earliest notation of the volcanic rocks which make up the bulk of the island. In 1890 Yates prepared a general descrip-

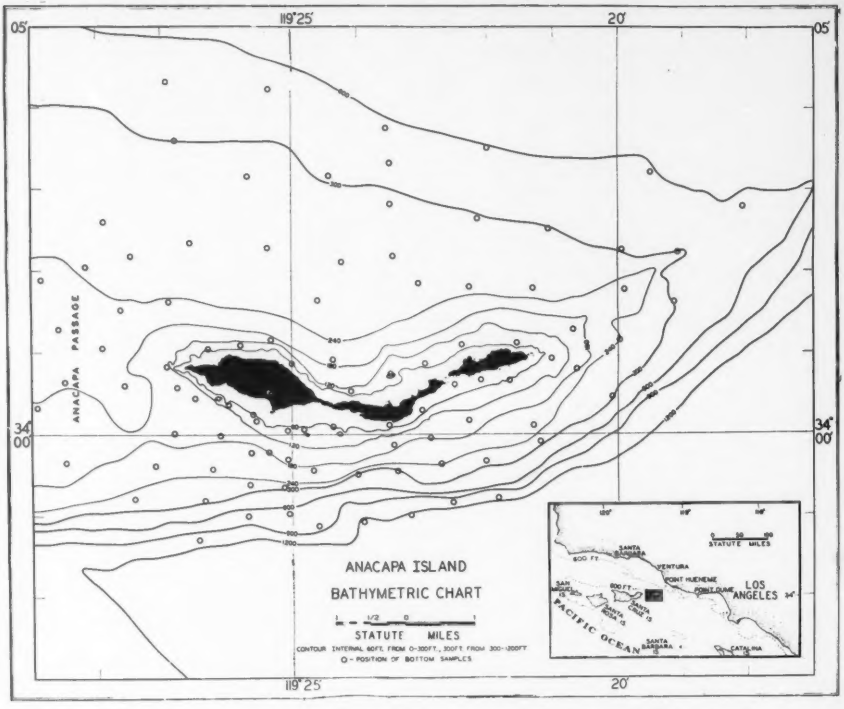

FIG. 1.—Location and bathymetric chart of Anacapa Island, California.

tion of the island's geology. His work was a standard reference for Anacapa until the appearance of a generalized columnar section by Kennett (1952) and of a more recent paper by Scholl (1959). Other previous knowledge of the submarine geology and shelf sediments is virtually limited to bottom notations on the U. S. Coast and Geoetic Survey Chart no. 5114 and Smooth Sheets nos. 5445 and 5446. Some generalizations of the shelf sediments were given by Trask (1931) and Revelle and Shepard (1939). Several detailed profiles of the Anacapa shelf have been constructed by Emery (1958).

A total of 98 Hayward, or orange-peel, grabs were collected from the shallow sea floor surrounding Anacapa Island during two cruises in the fall of 1957 and one cruise during the following spring. Approximately 70 of these samples are from the insular shelf which comprises an area of about 36 square miles. Six days were spent in mapping the geology of the island; most of the work was done during a three day sojourn in November 1957.

PHYSIOGRAPHY

Anacapa is a gently northward-dipping cuesta of volcanic rocks. The island is about 5 miles in length and is slightly arcuate (fig. 1). It is 3700 ft in maximum width and has an area of about 1 square mile. Two narrow gaps or passages divide the island into three distinct parts that are usually referred to as the eastern, central, and western islands. The western island is the largest of the three and rises to a summit elevation of 930 ft; the central and eastern islands have flat summits near an elevation of 250 ft.

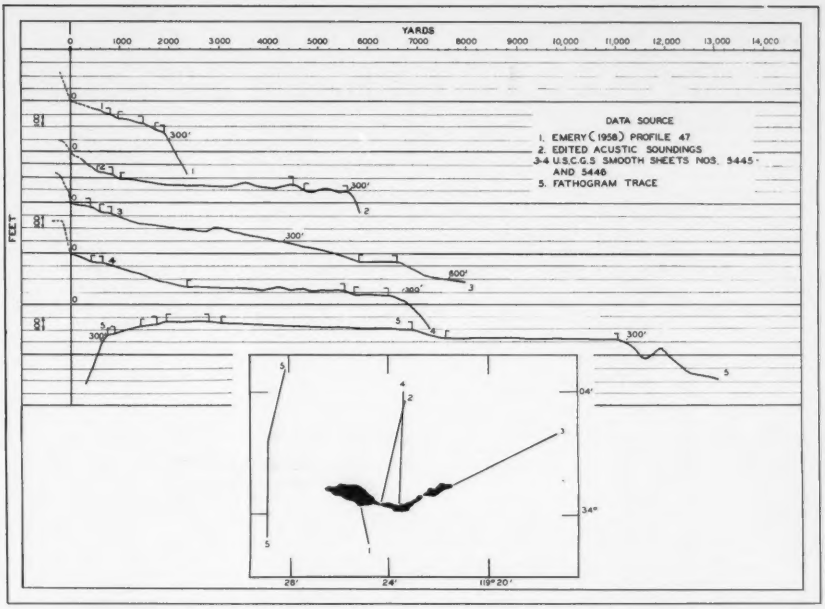

FIG. 2.—Submarine profiles of the Anacapa shelf. Symbols and] bracket inner and outer edges of terraces.

Drainage is principally to the north by intermittent resequent streams flowing down the dip slopes of tilted volcanic rocks. Streams are best developed on the higher western island, poorly developed on the central island, and practically non-existent on the eastern island. The annual precipitation is about 13 in (Dunkle, 1950).

Sea cliffs of spectacular height surround Anacapa Island. Along the southern coasts of the eastern, central, and western islands precipitous walls rise respectively to elevations of 250, 325, and 930 ft. Landsliding is very active on the cliff faces especially on the higher southern or anti-dip coastal cliffs. Fault- and joint-controlled sea caves are very numerous along the bases of the cliffs; more than 100 caves have been noted along the Anacapa shore. Anacapa has been greatly modified by incessant wave-attack since the beginning of the general post-Wisconsin rise in sea level. Evidence of this is clearly shown by the precipitous sea cliffs, the sea caves and between island passages, and the numerous outlying sea stacks which are found off all shores of the island.

Two wave-cut terraces can be recognized on the island. The highest terrace is on the north and west sides of the western island at an elevation of about 600 feet. The lower terrace is at an elevation near 250 feet and forms the flat summits of the central and eastern islands (fig. 1); it is also well developed on the western island. Raised wave-planed terraces indicate the island is not tectonically stable; thus it cannot be concluded that the island will be eventually wave-leveled. However, it is reasonable to conclude that if present conditions continue Anacapa may some day be charted as a bank or shoal rather than an island.

SUBMARINE TOPOGRAPHY

General

Knowledge of the sea floor topography surrounding Anacapa Island is primarily based on data taken from U. S. Coast and Geodetic Survey Smooth Sheets nos. 5445

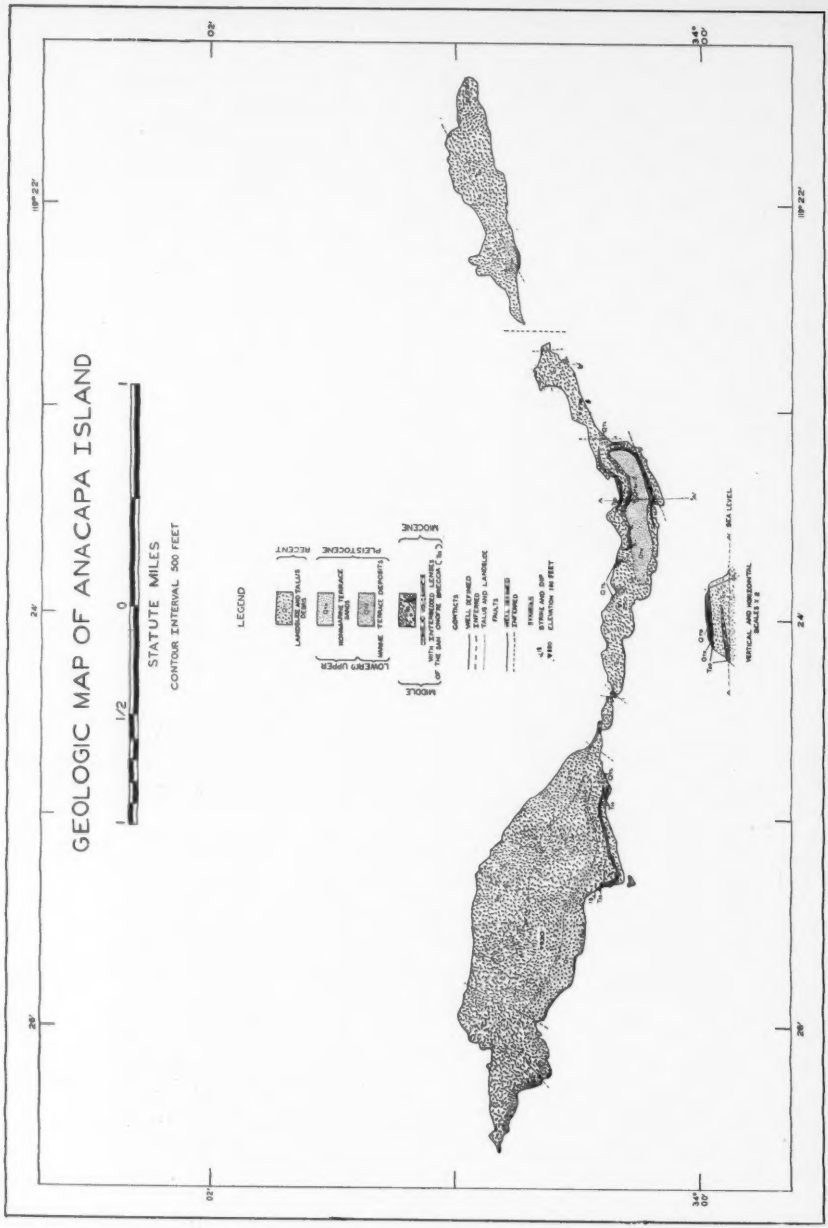

FIG. 3.—General geologic map of Anacapa Island.

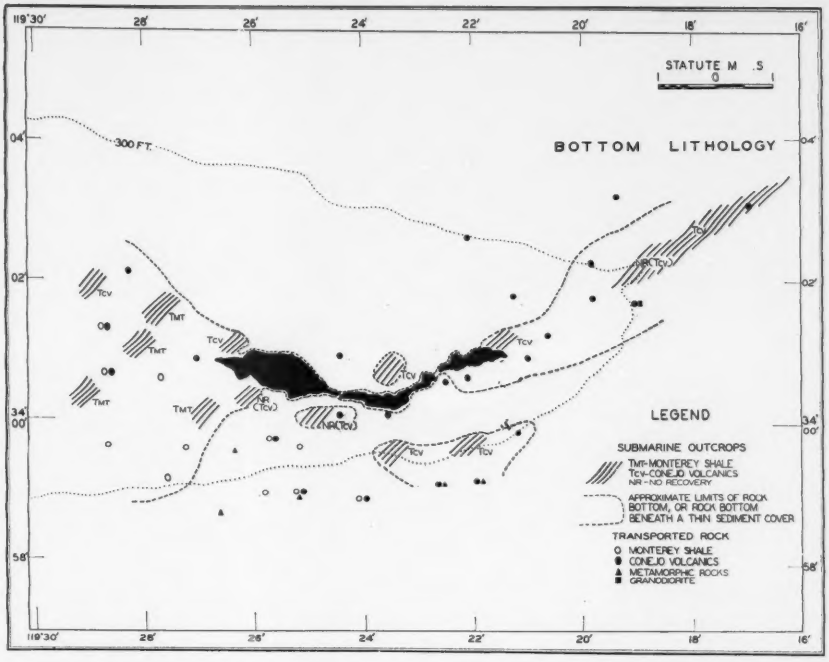

FIG. 3.—General geologic map of Anacapa Island.

FIG. 4.—Areas where submarine outcrops and transported rocks were located.

and 5446. These sheets were used to construct the general bathymetric chart shown on figure 1. Additional information was obtained from fathogram traces and from several detailed bottom profiles constructed by an edited acoustic sounding technique described by Emery (1958).

Nearly 14 of the 36 square miles of shelf area surrounding the island border its northern side. Less than 7 square miles flank the southern coast. The northern shelf at its maximum is about three miles wide; the southern shelf averages less than one-half this width. East of the island the shelf is intermediate in seaward extent between its northern and southern counterparts. To the west, the shelf extends beneath Anacapa Passage and merges with the eastern shelf of neighboring Santa Cruz Island, which lies about 4 miles from Anacapa (fig. 1).

A prominent oblique-slip fault is at the base of the southern submarine slope (Shepard and Emery, 1941, p. 47, and Corey, 1954). This fault is inferred from the steepness of the slope, which reaches 19 degrees, and its straightness. The proximity of the fault to the island is clearly responsible for the narrow southern shelf. Another fault is probably present at the base of the northern insular slope, but topographic evidence there is not entirely convincing; maximum gradients on this slope are about 6°.

Shelf Edge

Gradients exceeding 5 degrees mark the outer edge of the Anacapa shelf. The shelf edge, however, is not at uniform depths around the island. It is deepest to the north, where 13 measurements yield an average depth of 300 ft. The break in slope south of the island is somewhat shallower and averages 270 ft. East of the island the shelf edge is not well defined, but appears to be at a depth near 270 feet.

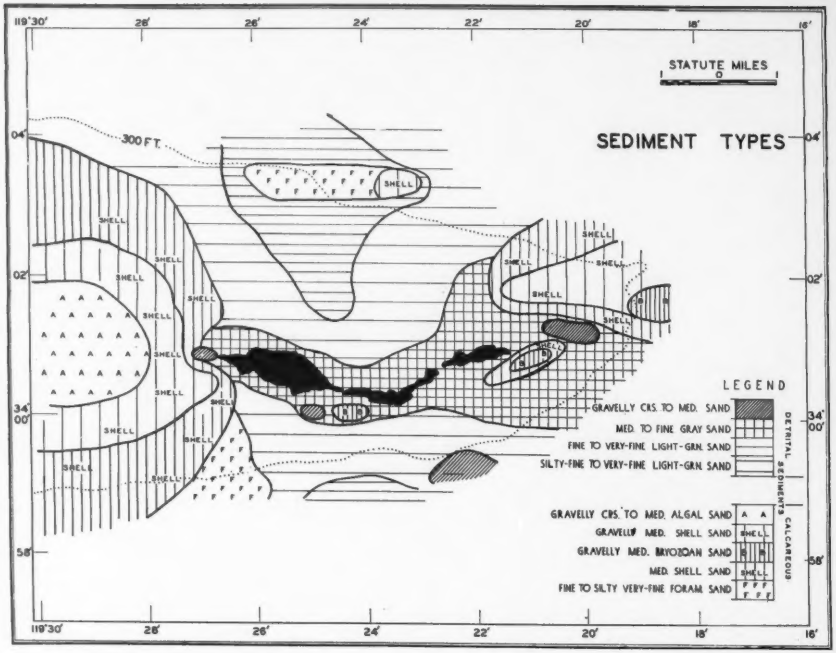

FIG. 5.—Distribution of sediment types; control stations for this figure and for figures 7 through 12 are shown on figure 1.

Shelves and Terraces

Steep submarine declivities ranging in slope from 2°30′ to more than 5 degrees flank the island near shore. The average declivity out to a depth of about 270 ft is 1°30′. Except for a few terraces, the sea floor south of the island slopes rather uniformly to the outer edge of the shelf. However, the nearshore inclination of the northern shelf is greatly reduced at a depth of about 270 ft, which usually occurs at a distance of 1 mile from the island. Beyond this depth a broad terrace with an over-all gradient of 0°05′ extends to near the shelf edge.

Four submarine terraces can be recognized on the southern shelf. The seaward edges of these flattenings are at depths of 60, 100, 180, and 240 ft (fig. 2). It is believed they are wave-cut features carved by lowered Wisconsin sea levels (Emery, 1958). North of the island the 60-ft terrace is present, but the corresponding 100-ft terrace is at a depth of 120 ft. Both of the deeper shelf flattenings are on the northern shelf, but the matching 240-ft terrace is fully 30 ft deeper. A 330-ft terrace typically forms the outer edge of the northern shelf (fig. 2, profiles 4 and 5).

The shelf area beneath Anacapa Passage is the westward continuation of the prominent 180-ft terrace. Examination of profiles (fig. 2, profile 5) reveals that the passage floor is inclined to the north. The direction of this declivity, and the facts that (1) the average shelf break is deeper north of the island and (2) submerged terraces are deepest on the northern shelf, are interpreted to mean that the shelf has been slightly tilted to the north during Recent time. The shelf itself is a wave-carved platform cut across north-dipping Middle Miocene rocks by fluctuating Pleistocene sea levels.

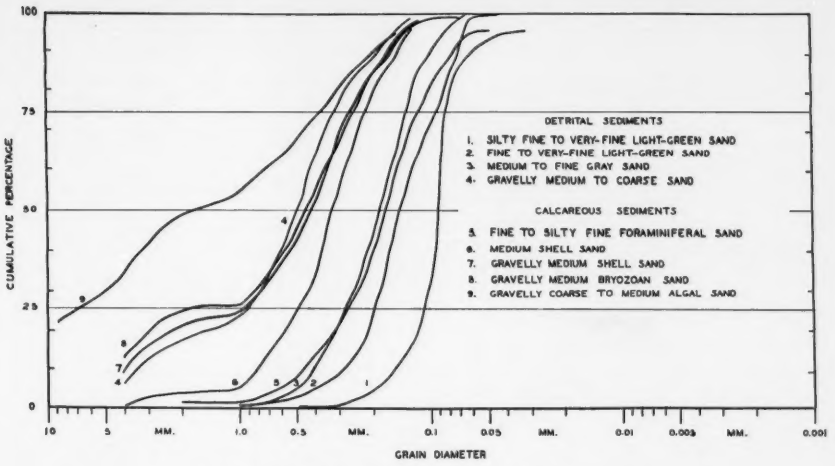

FIG. 6.—Typical cumulative curves of sediment types.

GEOLOGY

Stratigraphy of the Island

Conejo Volcanics (*Middle Miocene*).—Black to dark-red vesicular and porphyritic andesitic extrusive and pyroclastic rocks of the Conejo Volcanics of Middle Miocene age are the oldest and principal rocks on Anacapa Island (Scholl, 1959). These rocks consist of massive to thinly-bedded andesite lavas, autobrecciated flows, lenses of lapilli tuff, tuff breccia, and beds of volcanic breccia and agglomerate. The volcanic series strikes west or slightly north of west and dips a maximum of about 20 degrees north. A maximum thickness of about 1700 ft is exposed on the south side of the western island (fig. 3).

Phenocrysts of labradorite make up about 30 percent of the andesite lavas. The feldspar occurs as euhedral laths up to 5 mm in length. Phenocrysts of hypersthene and augite also occur, but they are less abundant and rarely exceed 1 mm in length. Pyrite is abundant as disseminated grains; secondary opal and chalcedony are present as vein and cavity fillings.

San Onofre Breccia (*Middle Miocene*).—Interbedded with the Conejo Volcanics are two strata of the San Onofre Breccia which are separated stratigraphically by 100 to 150 ft of volcanic rocks. Each interbed has a maximum thickness of about 35 to 40 ft. These breccias consist of blocks and angular cobbles and pebbles of glaucophane schist, hornblende schist, chloritic and talcose schist, and pink quartzite interbedded with pink and light-green silty sandstones composed of rock and mineral fragments common to the coarser breccia beds. Some of the metamorphic blocks are estimated to weigh more than 100 lbs.

Both units of San Onofre Breccia crop out near the eastern ends of the western and central islands. Exposures on the western island are only near the base of the precipitous southern sea cliffs, whereas outcrops are on both the northern and southern sides of the central island. The San Onofre Breccia interbeds are restricted to the basal 200–300 ft of the exposed volcanic series (Scholl, 1959).

Marine Terrace Sands (*Lower Pleistocene* (?)).—A poorly consolidated fossiliferous sand less than 3 ft thick crops out intermittently for a few hundred feet along the northern edge of the summit terrace atop the central island (fig. 3). The sand was doubtless laid down contemporaneously with the cutting of the summit terrace. It unconformably overlies the San Onofre Breccia from which most of its clasts have been derived.

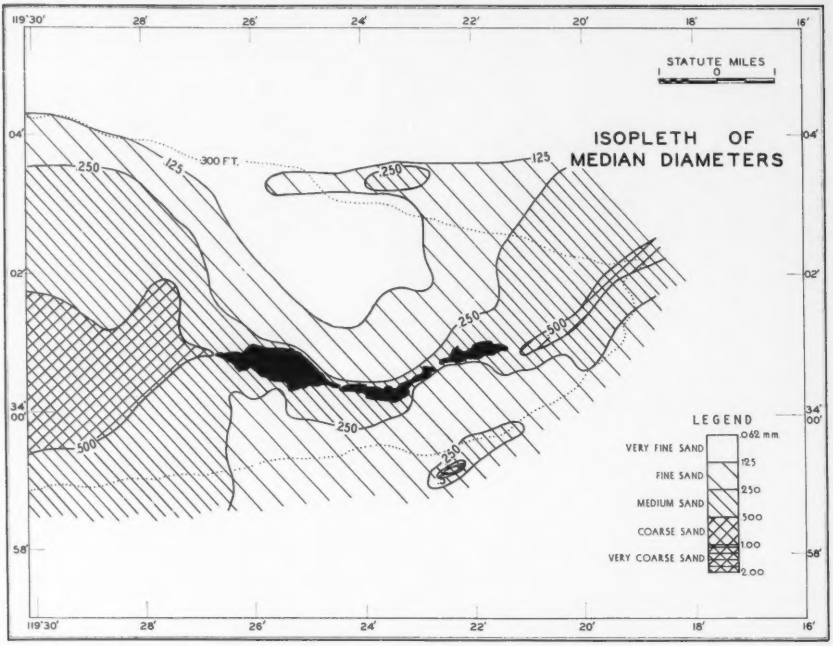

FIG. 7.—Isopleths of median diameters of shelf sediments.

TABLE 1.—*Average character of sediment types*

	Median diameter (mm)	Sorting coefficient	$CaCO_3$ %	Organic matter %	Heavy minerals %	Number of samples analyzed
Detrital sediments						
Gravelly medium to coarse sand	0.505	1.70	43.2	0.27	8.3	4
Medium to fine gray sand	0.185	1.51	16.1	0.48	8.7	14
Fine to very fine light-green sand	0.145	1.40	29.9	0.91	6.8	21
Silty fine to very fine light-green sand	0.091	1.16	14.2	1.15	3.5	2
Calcareous sediments						
Gravelly coarse to medium algal sand	1.700	2.80	85.6	1.30	7.0	2
Gravelly medium bryozoan sand	0.468	2.74	68.9	0.40	11.0	2
Gravelly medium shell sand	0.425	1.91	70.7	0.67	7.3	9
Medium shell sand	0.330	1.51	73.7	0.77	4.3	11
Fine to very fine foraminiferal sand	0.175	1.56	54.7	1.08	9.0	3

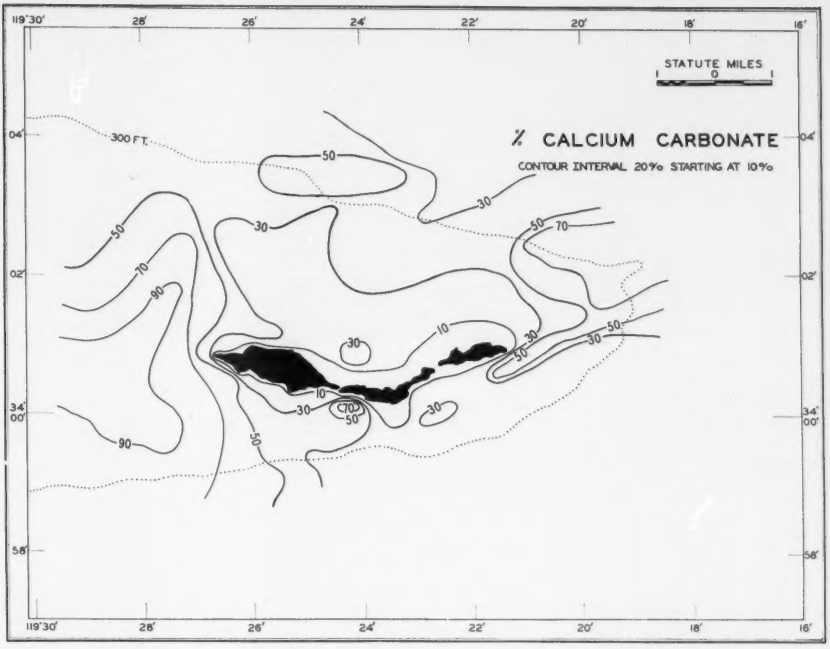

FIG. 8.—Distribution of calcium carbonate in shelf sediments.

Mollusks collected from the sand indicate a time correlation which the Pleistocene of California (Grant and Gale, 1931). Foraminifera suggest a time correlation with the Lower Pleistocene San Pedro Formation of the Los Angeles Basin, particularly with its shallow-water biofacies. The abundance of two warm water species, *Elphidium fax*, and *E. microgranulosum*, are thought to signify an interglacial deposit probably older than Sangamon (Dr. O. L. Bandy, personal communication).

Nonmarine Terrace Sands (Upper Pleistocene to Recent).—The northern edge of the summit terrace near the eastern end of the central island is overlain by about 25 to 30 ft

TABLE 2.—*Gross character of sediments*

	Eastern shelf	Western shelf	Southern shelf	Northern shelf	Entire shelf
Median diameter (mm)	0.401	0.714	0.181	0.142	0.360
Sorting coefficient	2.34	1.84	1.32	1.42	1.73
Carbonate content (%)	60.5	75.0	22.3	25.5	45.4
Dominant bioclast	shell then bryozoan	shell then algae	shell then Foraminifera	Foraminifera then shell	shell then Foraminifera
Light mineral and rock particle content (%)	30.8	18.7	67.5	67.5	46.3
Heavy mineral content (%)	8.1	5.6	9.5	6.1	7.6
Organic matter content (%)	0.6	0.7	0.7	0.9	0.7
Number of samples analyzed	10	11	16	13	50

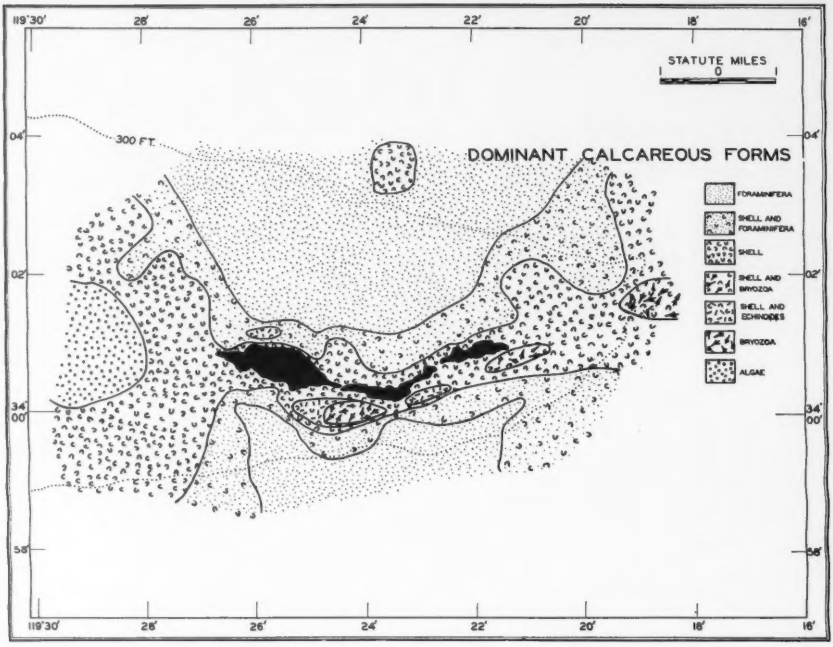

FIG. 9.—Distribution of dominant calcareous forms in shelf sediments.

of alluvial sands and silty sands which thin rapidly to the south and are only a few feet thick at the southern edge of the terrace. The thinning to the south and the fact that the mineralogy of the sands is nearly identical with that of the underlying San Onofre Breccia indicate that the San Onofre Breccia cropped out on a higher part of the island which formerly existed to the south. It is evident that the summit terraces of the eastern and central islands are remnants of a former fringing terrace which skirted the northern side of a larger and higher island of Pleistocene age.

Shelf Stratigraphy

Rocks recovered from the Anacapa shelf were classified as either transported debris, or as rock indicating the close proximity of submarine outcrops. The numerous criteria given by Emery and Shepard (1945, p. 484) were used in making the distinction. Transported rocks are not considered here because they cannot reliably indicate the shelf stratigraphy.

In all, fourteen submarine outcrops were located (fig. 4). The lithologic character of most of these can readily be correlated with formations cropping out on Santa Cruz and Anacapa Islands. Exposed rock bottom is most widespread over the western shelf area beneath Anacapa Passage; however, parts of the passage floor are covered by a thin veneer of Recent and relict Pleistocene sediments. Exposed and partly buried outcrops are also abundant on the eastern and southern shelves, but are generally lacking north of the island except in shallow nearshore areas.

Conejo Volcanics (Middle Miocene).—Outcrops of the volcanic series are in many areas in shallow water off all shores of the island. East of the island a nearly continuous outcrop ridge is traceable beyond the

TABLE 3.—*Mineralogical composition of shelf sediments*

	Eastern shelf							
	Average of 10 samples		Western shelf		Southern shelf		Northern shelf	
	(A) % in non-calcareous fraction	(B) % in total sample	(A)	(B)	(A)	(B)	(A)	(B)
Light minerals								
labradorite	40.6	16.0	40.9	10.2	46.3	38.0	47.8	35.7
untwinned albite	p	p	p	p	p	p	p	p
orthoclase	0.5	0.2	0.4	0.1	0.1	0.1	0.8	0.6
quartz	3.5	1.4	4.9	1.2	3.1	2.5	2.8	2.1
glauconite	3.2	0.9	2.4	0.6	1.4	1.1	1.7	1.2
rock particles	31.1	12.2	25.8	6.4	35.4	27.4	35.3	26.3
opal (sponge spicules)	p	p	0.4	0.1	0.5	0.4	3.1	1.5
biotite	p	p	p	p	p	p	p	p
Total	78.1	30.8	74.9	18.7	86.8	67.5	90.6	67.5
Heavy minerals								
augite	12.0	4.7	13.5	3.3	7.0	5.5	5.5	4.0
hypersthene	7.3	2.9	6.2	1.5	4.0	3.1	2.1	1.6
pyrite	0.2	p	0.4	0.1	0.4	0.3	0.3	0.2
limonite	0.2	p	1.00	0.2	0.3	0.2	0.1	0.1
rock particles	0.3	0.1	0.2	p	p	p	0.1	p
glaucophane	0.2	p	0.3	p	0.1	0.1	p	p
crossite	—	—	p	—	p	p	p	p
epidote	p	p	p	p	p	p	p	p
hornblende	p	p	0.3	p	p	p	p	p
colorless garnet	p	p	p	p	p	p	p	p
zircon	—	—	—	—	p	p	p	p
ilmenite and magnetite	p	p	0.3	p	p	p	p	p
Total	20.5	8.1	22.4	5.6	12.2	9.5	8.2	6.1
Calcium carbonate	—	60.5	—	75.0	—	22.3	—	25.2
Organic matter	1.4	0.6	2.7	0.7	1.0	0.7	1.2	0.9
Sum of totals	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0

p=less than 0.1 percent.

shelf edge toward the mainland. Other exposures are at or near the outer edge of the southern shelf and over parts of the western shelf.

Monterey Shale (Middle Miocene).—Cream-colored calcareous and siliceous shale and gray mudstone crop out on the shelf area beneath Anacapa Passage (fig. 5). The shales closely match Bremner's description of the Monterey Formation which crops out near the eastern end of neighboring Santa Cruz Island (Bremner, 1932, p. 32). On Santa Cruz Island rocks of the Monterey Formation stratigraphically overlie the Conejo Volcanics. Beneath Anacapa Passage the shales crop out at least 180 ft lower than the 1700 ft of volcanic rocks exposed on the western island. This may indicate that the two formations are in fault contact beneath Anacapa Passage. However, on Santa Cruz Island shales of the Monterey Formation are known to be interbedded with the Conejo Volcanics; therefore, faulting need not be relied upon to explain outcrops of the Monterey Formation on the passage floor.

Structure

Anacapa lends its name to the structural province of Anacapia, which comprises the Santa Barbara Islands and the Santa Monica Mountains on the mainland (Reed and

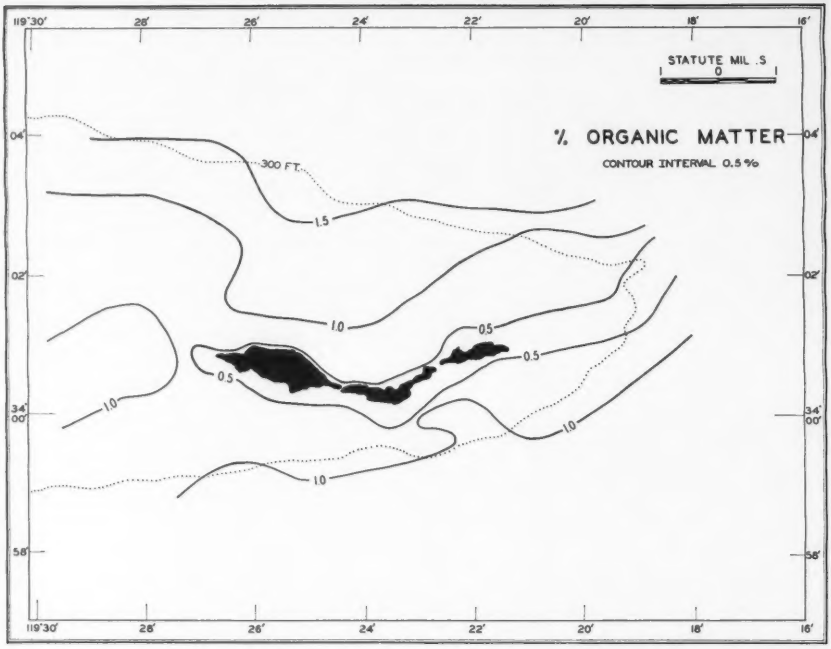

FIG. 10.—Distribution of organic matter in shelf sediments.

Hollister, 1936, p. 105). Anacapa and the other Santa Barbara Islands rise above a submerged anticlinal ridge which extends seaward from the Santa Monica Mountains. Anacapa itself is a gentle monocline of north-dipping rocks. A maximum dip of about 20 degrees occurs on the western island, those of the central and eastern islands are typically less than 5 degrees.

SHELF SEDIMENTS

Analyses

Textural parameters of shelf sands were determined with the Emery Settling Tube (Emery, 1938). Finer grained sediments were analyzed by the pipette method (Krumbein and Pettijohn, 1938, p. 162). Samples containing important gravel, sand, and silt-clay fractions were screened to separate these components.

Mineral determinations were facilitated by separating light and heavy (greater than 2.96) specific gravity minerals after treatment with cold dilute hydrochloric acid. The light mineral suite was stained to distinguish quartz and sodic and potash feldspars (Twenhofel and Tyler, 1941, p. 131).

Chemical analyses were made to determine calcium carbonate and organic carbon. Carbonate content was measured by leaching a known weight of sediment with dilute hydrochloric acid, which, although reliable for comparative purposes, tends to give values from 2 to 5 percent too high. Organic carbon was determined semiquantitatively by the potassium dichromate reduction technique described by Allison (1935). Organic carbon percentages were multiplied by a factor of 1.7 to convert to total organic matter (Emery and Rittenberg, 1952 ,p. 781).

Sediment Types

The Anacapa shelf deposits can be grouped into calcareous and detrital sediments. Calcareous deposits are defined as

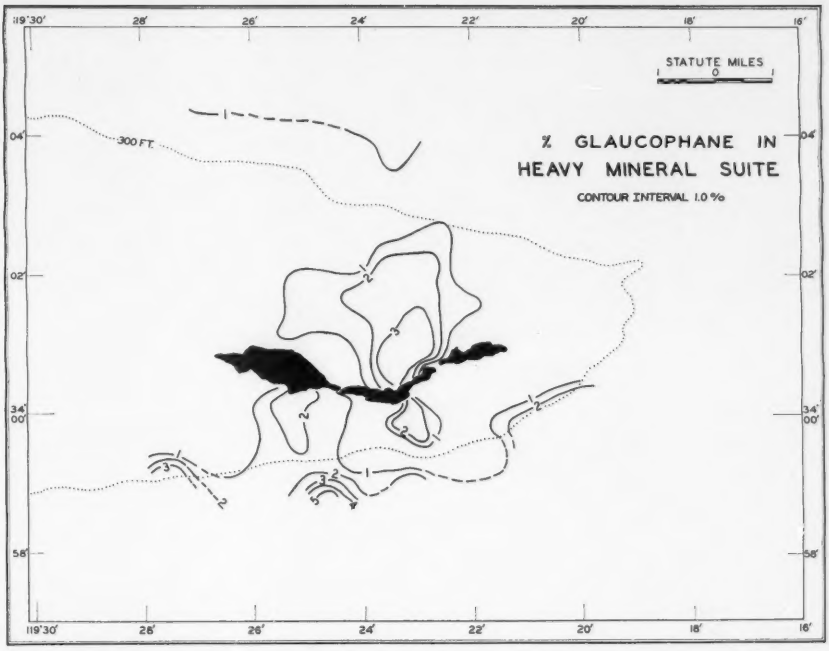

FIG. 11.—Distribution of glaucophane in heavy mineral suite of shelf sediments.

having carbonate contents greater than 50 percent. Detrital sediments are divided into sediment types based on their median diameter (for example gravel, sand, silt), on secondary grade classes which compose more than 10 percent of the deposit (for example gravelly, sandy, silty), and on their color (wet). Calcareous accumulations are classified in a similar manner, except that the dominant calcareous organism is used to further differentiate the sediment type.

Altogether nine sediment types can be distinguished, four of which belong to the detrital group. Detrital sediments are primarily on the northern and southern shelves and are the most widespread of the insular deposits; they cover approximately 57 percent of the shelf platform or about 20 square miles. Conversely, calcareous sediments are largely restricted to the eastern and western shelves and overlie the remaining 16 square miles of shelf areas. The distribution of the sediment types is shown on figure 5. Their textural, mineralogical, and chemical properties are listed in table 1, and their average cumulative curves are given on figure 6.

Sedimentary Environments

Two distinct sedimentary environments are delimited by the distribution and character of the sediment types (fig. 5). The first of these is one of little or no detrital sedimentation; this condition exists over the insular platform east and west of the island. In marked contrast, the second depositional environment is one of detrital sedimentation and is prevalent north and south of the island. The physical and chemical properties of the sediments forming under these sedimentary conditions are compared in table 2.

Eastern and Western Shelves.—Shelf sediments east and west of the island are well-sorted medium- to coarse-grained highly calcareous sands owing to the accumulation of animal and plant fragments (figs. 7, 8, 9,

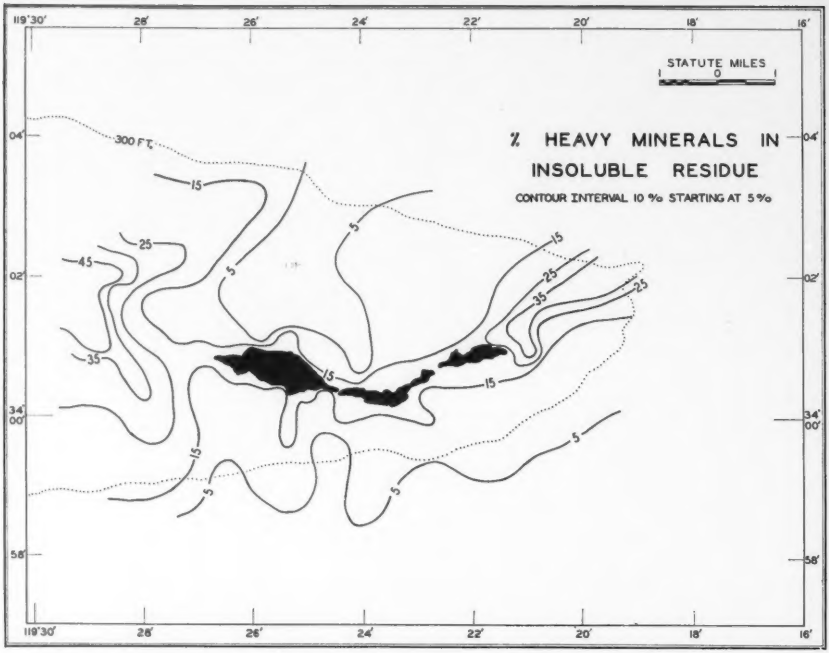

FIG. 12.—Distribution of heavy minerals in non-calcareous fraction of shelf sediments.

and table 2). That these two areas are characterized by slow sedimentation is shown by the fact that glauconite, an authigenic mineral which forms in areas of slow or insignificant detrital sedimentation (Emery, 1952), is most abundant in the sediments of the eastern and western shelves (table 3).

Coarse- and medium-grained detrital sands are also on the eastern and western shelves, but these sediments are closely associated with submarine outcrops and represent residual and relict accumulations. Rounded pebbles and cobbles showing evidence of transportation are widespread over the two shelf areas, but these rocks are also closely associated with submarine outcrops (fig. 4). It is reasonable that the transported rocks are lag deposits dispersed during lowered Pleistocene sea levels. In one instance, however, a granodiorite cobble was recovered near the edge of the eastern shelf which must have been rafted into place (Emery, 1955) (fig. 4). The nearest outcrop of granitic rock is about 20 miles to the west on Santa Cruz Island (Bremner, 1932).

The sedimentary environments of the eastern and western shelves are not conducive to the deposition nor the preservation of organic detritus; deposits east of the island have the lowest average organic matter content (0.6 percent) of the four major shelf areas (table 2). However, because plant tissue is present in algal sediments beneath Anacapa Passage, sediments on the western shelf have organic matter contents greater than 1.0 percent (fig. 10 and table 1).

A combination of several factors is responsible for the paucity of Recent detrital sedimentation on the eastern and western shelves. These are: (1) both areas lie seaward of the tapering ends of the island and consequently little sediment is contributed directly to them; (2) both shelf areas are exposed to currents which sweep across the Anacapa shelf from the northwest or south-

east; these currents are probably strong enough to prevent the deposition of the finer (silt-clay) detrital clasts; and (3) tidal currents, in particular over the western shelf beneath Anacapa Passage, are probably competent enough to prevent the accumulation of most detrital clasts.

Rock bottom is exposed over much of the shelf area east and west of the island and most of the associated calcareous sediments show clear signs of abrasion and partial solution. It is apparent that the sedimentary blanket is thin on these two shelf areas and that some of the calcareous clasts are relict accumulations of Late Pleistocene age.

Northern and Southern Shelves.—North and south of the island the shelf deposits consist of well-sorted coarse to very fine sands which are principally composed of low specific gravity mineral and rock fragments (figs. 5, 6, 7, and table 2). A consistent seaward decrease in grain size is interpreted to mean that these clasts are largely derived from the island. Reversals in this trend occur, but they can be correlated with known submarine outcrops or with localized accumulations of coarse calcareous sediments (figs. 4 and 5). Cobbles and pebbles showing evidence of transportation are principally fragments of the Conejo Volcanics and are closely associated with submarine outcrops of the formation. However, rounded shale fragments from the Monterey Formation are present south of the island where outcrops of the formation are not known. Transported rock is rare north of the island and apparently Pleistocene lag deposits on the northern shelf have been buried by Recent sedimentation.

Studies of the distribution of median diameters and of glaucophane, a highly pleochroic sodic amphibole which is derived only from outcrops of the San Onofre Breccia and which is present in the shelf sediments as grains commonly between 0.250 and 0.062 mm, indicate that detrital clasts finer than about 0.125 mm tend to bypass the shelf area and are deposited on the flanking insular slopes and in adjacent deep sedimentary basins (figs. 7 and 11). The silty very fine sands (fig. 5) and the spreading of the isopercentage lines of glaucophane on the broad 270-ft terrace north of the island suggest that by-passing is not as important for this shelf as it may be for the opposite southern shelf which lacks both the outer terrace and the very fine sand.

Heavy minerals constitute a lower percentage of the non-calcareous fraction of the finer grained sediments north and south of the island than they do of the coarser deposits overlying the eastern and western shelves. This is due to the relatively large size of augite and hypersthene grains which commonly form as much as 90 percent of the heavy mineral suite. Within the sediments these two minerals are commonly larger than 0.250 mm, and thus they tend to accumulate in the coarser sediments east and west of the island. In agreement with this tendency to concentrate in the coarser shelf sediments, the heavy mineral content of the non-calcareous fraction decreases with distance north and south from the island, and increases with distance to the east and west (fig. 12).

It is evident from an increasing carbonate content that detrital sedimentation becomes less prevalent with distance from shore (fig. 8). The increasing calcareous matter is chiefly caused by an increasing abundance of Recent foraminiferal tests. It is of interest to note here that, whereas some of the calcareous clasts east and west of the island are much abraded and are of Late Pleistocene age, most of the calcareous content in the sediments of the northern and southern shelves is of Recent deposition.

Organic matter also increases with distance from shore (fig. 10), apparently in response to the seaward decreasing sediment grain size (Trask, 1931). Owing to oxidizing conditions and slow sedimentation rates, organic detritus is not readily preserved in the sediments and averages only 0.8 percent for the two shelf areas. This value is considerably lower than the 2.50 percent determined by Trask for many of the world's nearshore sediments (Trask, 1939, p. 429).

Active cliff erosion is thought to contribute the greatest volume of detrital clasts to the northern and southern shelves. Streams must also contribute appreciable amounts of sediment, but probably because of dispersing currents shelf deposits do not noticeably differ adjacent to stream mouths. The rate of sedimentation during Recent time has probably been highest north of the island

since the unconsolidated sediment cover there is thickest; this is judged from the general uniform nature of the bottom deposits and from the paucity of lag deposits and submarine outcrops on the northern shelf.

SUMMARY

It has been shown that the relationship, character, and distribution of the unconsolidated shelf sediments of Anacapa Island reflect the geology of the island and the shelf and the parameters of the surrounding sedimentary environments. The sedimentary environments also reflect the submarine topography and the relative position of the various parts of the shelf to the location of the island and to the pattern of prominent shelf currents. The geology of the island and the combined and relative affects of these several factors and the nature of shelf sediments are summarized below:

1.—Anacapa Island is approximately 5 miles long, 3700 ft wide, and consists of three small islands; precipitous sea cliffs surround each island. A wave-planed summit terrace at an elevation of approximately 250 ft is atop the eastern and central islands; an additional terrace is near an elevation of 600 ft on the western island which rises to a summit peak at 930 ft.

The island consists of a maximum of about 1700 ft of gently north-dipping andesitic flows and pyroclastic rocks of the Conejo Volcanics of Middle Miocene age. Two strata of San Onofre Breccia are interbedded near the base of the volcanic series. Marine and non-marine sands of Pleistocene age overlie the summit terrace on the central island. Submarine outcrops of the Conejo Volcanics are abundant on the shallow shelf east, west, and south of the island; submarine outcrops of the Monterey Shale (Middle Miocene) are west of the island beneath Anacapa Passage.

2.—East and west of the island the character of the shelf deposits is largely controlled by sweeping currents which prevent the accumulation of detrital clasts and the finer calcareous bioclasts. As a consequence, most of the deposits are coarse-grained sands and gravels of calcareous animal and plant fragments. These organic remains have been accumulating since the Late Pleistocene. Deposits of coarse-grained relict and residual detrital sediments flank outcrops of the Conejo Volcanics and the Monterey Shale; thus, exposed bedrock locally affects the character and distribution of the sediment types.

3.—On the narrow and rather steep-sloping southern shelf the sediments are mainly medium- to fine-grained detrital sands. Most of these clasts are mineral and rock fragments derived from the Conejo Volcanics which make up the bulk of the island and which underlie and crop out over much of the shelf area. Rock and mineral clasts from the San Onofre Breccia form an instructive but minor portion of the deposits. Calcareous bioclasts are also an mportant part of the southern shelf sediments; foraminiferal tests of Recent age are most abundant over the outer two-thirds of the shelf, but in nearshore areas shell (mollusk), bryozoan, and echinoid bioclasts are most common.

The distribution and character of the southern shelf sediments are controlled primarily by shelf currents and submarine topography. Slopes exceeding 1°30′ near the outer edge of the shelf allow detrital clasts finer than approximately 0.125 mm to be swept beyond the shelf area. Submarine outcrops locally control the distribution and character of the bottom deposits and are typically flanked by relict and residual accumulations. Because of slow sedimentation rates and oxidizing conditions, particulate organic detritus is not readily preserved in any of the insular deposits and averages only 0.7 percent; there is, however, a relative concentration of organic matter in the finer grained shelf sediments.

4.—The shelf deposits of the northern shelf are comparable in most respects to those of the opposite southern shelf, and thus similar forces must shape the character and distribution of the insular shelf deposits north and south of the island. However, the gentle slope (0°05′) of the outer 270-ft terrace north of the island prevents the finer sand clasts from moving across the shelf to the adjacent insular slope and sedimentary basins. By-passing is clearly not as prevalent north of the island as it is to the south. Primarily for this reason the unconsolidated sedimentary blanket is thickest

north of the island and the rate of Recent sedimentation has been highest here.

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FIGS. 1–2, MARCH, 1960

WEATHERING OF THE GREAT PYRAMID[1]

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ABSTRACT

The Great Pyramid of Giza, Egypt, was built of limestones and limy sandstones of varying resistances to weathering. As a result, the blocks exhibit several different surface textures. When the facing was removed approximately 1000 years ago, products of weathering formed talus slopes that overflowed the tiers of exposed blocks and banked against the base of the pyramid. An estimate of the rate of formation of talus indicates that the pyramid should remain standing at least 100,000 years unless its destruction is sped by man's activities.

INTRODUCTION

During visits to the Great Pyramid of Khufu (Cheops) at Giza, Egypt, in February 1959, the writer was impressed by its extensive weathering. Originally, the pyramid was covered by a facing of smooth well-fitted limestone blocks, as described by Herodotus, Strabo, Pliny, and other early travelers. Part of the lowest course of facing still remaining at the middle of the north side has joints less than 1 mm wide between blocks as long as 1.5 m, presumably typical of the entire original facing. Because of the tight joints, steep (52°) slope, and dense nature of this limestone facing, the inner layers of core rock were doubtlessly well protected from rain water during most of the thousands of years since construction of the pyramid about 2800 B. C. When the facing was quarried away beginning in the ninth century to build mosques and other large and small structures at Giza and Cairo, the core rock became exposed to weathering. The nature of this weathering forms one of the interesting geological problems associated with the pyramid (Wagner, 1935), as distinct from the mystical ideas about the pyramid which were developed during the nineteenth century (Muck, 1958).

Appreciation is due to the Guggenheim Foundation for providing a grant for travel which made possible the visit to Giza.

LITHOLOGY

Blocks that form the core of the pyramid range from about $1.5 \times 1.5 \times 2.0$ m at the base to about $0.5 \times 1.0 \times 1.0$ m near the top. Most of at least the outer blocks were placed so that bedding planes are horizontal. They consist of four main varieties of rock: gray hard dense limestone, gray soft limestone, gray shaly limestone, and yellow limy shaly sandstone. Gradations between these main types are present. Some of the blocks of hard limestone are nummulitic like that of the Eocene bedrock of the promontory atop which the pyramid was built, but most are unfossiliferous. The other blocks of softer rock probably also came from underlying strata obtained at nearby quarries still remaining on all sides of the pyramid (Reisner, 1942, p. 10–12). The facing was of a different hard, fine-grained, white limestone which was obtained from quarries in Jebel Mokhattam 20 km distant and across the Nile River (Watson, 1911).

WEATHERING OF THE ROCKS

The four main kinds of rock have weathered differently. So little affected is the gray hard dense limestone that many blocks of it retain marks made by the quarry tools (fig. 1A). The builders evidently recognized the durability of this kind of limestone because they concentrated it at the corners, top, base, and openings of the pyramid. Locally, however, other concentrations occur at several levels with no obvious purpose. It is notable that the third pyramid at Giza (of Men-kū-Rē or Mycerinus), the smallest of the three, consists almost entirely of this kind of rock. More abundant than the hard limestone is the gray soft one; its weathering is characterized by pits that average about 1 cm in diameter and 2 cm in depth (fig. 1A). Many pits have a greater diameter at

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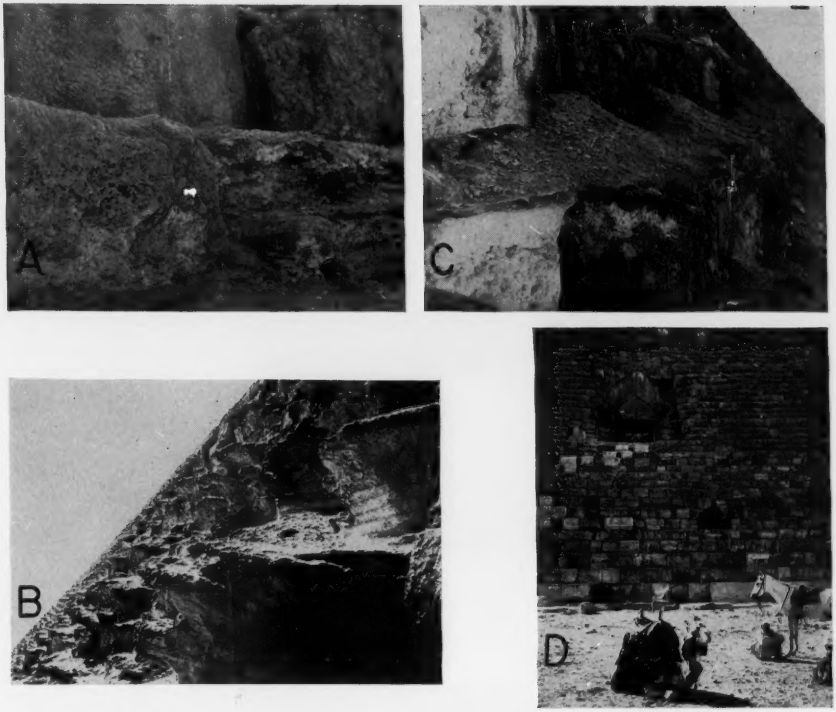

FIG. 1.—Details of Great Pyramid.
A. Lithology of core rock approximately 20 m above base at east side of pyramid. Block at upper left is gray hard dense limestone showing quarry tool marks; block at lower right is yellow limy shaly limestone partly broken down into small joint fragments; other two blocks are soft gray limestone having pitted surfaces.
B. General view of east side of pyramid approximately 60 m above base. Foreground blocks are of soft gray limestone in which niche formation has undermined top surface.
C. Talus mantling tiers of blocks approximately 40 m above base on north side of pyramid.
D. Entrance near middle of north side. Note lighter color and sharper edges of blocks below level of entrance than above it; these blocks were protected from weathering by cover of talus. White blocks at base behind horse are remnant of original facing of Tura Limestone; white blocks higher above base are concrete reinforcements.

depth than near the surface, as though the blocks had become somewhat case-hardened after being cut.

The other two chief rock types are less common than the two foregoing ones, a fortunate fact because of their much faster rate of weathering. The more common of the two is the gray shaly limestone, which on weathering has developed niches, many of them exceeding 20 cm in depth. The niches are rounded in form and have soft scaly walls. Many blocks have been undermined by niche development (fig. 1B) so that an apparently sound top surface is unsupported and likely to break away under the feet of of an unwary climber. It is not clear whether the 5- to 15-cm thick top surface is stronger than the lower part of the blocks because of case-hardening or whether the builders intentionally put the strongest side of the blocks uppermost. Lastly, there are scattered blocks composed of friable yellow limy shaly sandstone. This rock exhibits close parallel and intersecting joints on weather-

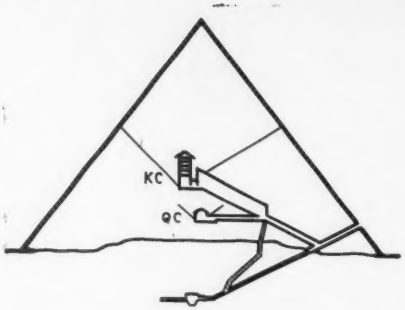

FIG. 2.—Internal arrangement in Great Pyramid, redrawn from Edwards (1947, fig. 14). Symbols KC and QC indicate so-called King's Chamber and Queen's Chamber, respectively. North is at right-hand side. At present, the pyramid is dry inside, even in the passage and chamber beneath the ground level.

ing which cause it to turn easily into rubble (fig. 1A). Typically, the weathered surface slopes in the same direction as the pyramid face, but less steeply. The surface is one on which a climber may easily slip, especially where many small joint fragments litter the surface. Most of the core blocks of the second pyramid (of Khaef-Rē or Chephren) are of this rock type, and as a result its present surface is loose and nearly unclimbable.

Weathering is restricted almost exclusively to the top and front faces of the blocks. The sides where they adjoin other blocks are usually unaffected. In fact, most of the joints are filled with gravelly to sandy to silty reddish sediment, some of which is probably an ancient mortar used to fill the gaps between the rough-hewn blocks. Exceptionally, the gaps between blocks are as much as 30 cm wide although most are only about 2 cm. The joint filling is less resistant to weathering than are the blocks themselves, as shown by the fact that the joints are commonly indented a few cm behind the faces of adjacent blocks. One additional minor aspect of the weathered surfaces is a brown surface film that is well developed on the south side of the pyramid, locally present on the east and west sides, and absent on the north side. It appears to be an early stage of desert varnish, and analyses reported by Lucas (1938) indicate a slight concentration of iron, manganese, silicon, and other elements in it.

Notably absent on the blocks is evidence of erosion by running water or wind. The former left no channels and the latter, no polished surfaces. In fact, sands sorted well enough to have been wind-blown were not found, and according to the nature of the nearby debris dump little was ever present on the pyramid. Unlike the pyramids at Chichinitza and Uxmal in Yucatan and at Angkor Wat in Cambodia, the Egyptian pyramids have not been attacked by the wedging of roots, because the rainfall is too low to permit growth of large plants.

PRODUCTS OF WEATHERING

The products of weathering reveal almost as much about the process as the remaining weathered blocks themselves. Some of the more soluble constituents may have been carried away in solution. Some were deposited in joints, probably as calcium carbonate cement and certainly as gypsum veins, several of which were observed both outside and inside the pyramid. At present, the pyramid is dry inside, even in the passage and chamber beneath the ground level (fig. 2). Measurements by Sutton (1945) show a humidity within the pyramid of only 47 to 61 percent, part of which is contributed by the perspiration of tourists. A 1 to 3 mm thick coating on the blocks was noted on the walls of the so-called Queen's Chamber and in the horizontal passage leading to it. None was found elsewhere. This material consists of two to eight layers of salt of the following composition:

Sodium	35.22%
Potassium	.02
Calcium and magnesium	.40
Chlorine	54.14
Sulfate	1.01
Water-insoluble	9.21
	100.00%

The water-soluble portion is thus about 99 percent halite, NaCl. Presumably, the salt was deposited during the few years that elapsed after the chamber was completed and before the pyramid was finished and faced. During this period the top was probably relatively flat (Lauer, 1952) and able to collect rain water which could seep through the joints, dissolving some salts in the rock for later deposition in the open space of the

chambers. The absence of the coating in the King's Chamber may be a result of the fact that the pyramid was finished so soon after this chamber was built that there was no time for leaching and deposition of salt, or that the salt was derived only from blocks that immediately line the chamber walls. In the King's Chamber these blocks are of granite, not limestone.

During the process of weathering the bulk of the rock remained as talus composed of angular gravel-sized fragments in a matrix of poorly sorted sand and silt. This talus is present on top of each tier of blocks on the north and south sides of the pyramid where it reaches a thickness of 30 cm and a width of 100 cm, although it is generally thinner and narrower, depending upon the width of the underlying blocks (fig. 1C). Its looseness and sloping surface, as well as its concealment of the soundness of the underlying blocks, makes the north and south sides extremely dangerous to walk upon. My guide, the pyramid climber Hafnowi, had once followed a tier across part of the north side only by shoveling aside the debris ahead of him in order to rescue a stranded boy. On the west and east sides the debris is much thinner (fig. 1A), having been removed in 1956 and 1957 for the filming of a movie. Nevertheless, the thickness is commonly about 2 cm, concentrated at the back middle parts of the blocks. In these areas it is notable that the lithology of the debris closely corresponds to that of the block immediately above and behind, and little talus fronts the blocks of hard limestone.

During the thousand years or so since the facing was stripped away, talus overflowed the restricted areas on top of the tiers of blocks so that much of it found its way down to the base of the pyramid. On each of the four faces of the pyramid, a lighter colored and lesser weathered area of blocks (fig. 1D) marks a part that was covered and protected by talus until about 1940 (Reisner, 1942, pls. 1, 2, 26; see also, the excellent camera lucida drawings made by E. W. Lane in 1826 and published for the first time by Ceram in 1957). The apex of the debris cone on each face reached about 18 m above the base, and its top curved down to the corners of the pyramid. Its slope should have been approximately the same as that of the talus still lying against the bases of the second and third pyramids, 30°. Although the talus doubtlessly contained some fallen blocks, most of it must have been identical with the fragments of weathering still present on top of the tiers of blocks. In fact, such is the composition of the basal portion of a talus cone still present at the west side of the Great Pyramid. Computation of that part of the volume of a cone, having a slope of 30° and a diameter equal to the diagonal of the pyramid, that lies outside the pyramid (52° side slope, 230-m base, 146-m height—Edwards, 1947, p. 87) shows that the volume of the talus must have been about 49,000 cubic meters. An additional 1000 cubic meters cover the north and south sides now and covered the east and west sides prior to 1956. This total of 50,000 cubic meters would have required a rate of formation of approximately 50 cubic meters per year since weathering started about 1000 years ago. It corresponds to an average loss of about 0.2 mm per year over the whole surface of the pyramid, allowing a 50 percent porosity of the talus. However, it annually represents only about 0.01 percent of the total volume—showing that the pyramid, while not eternal, may remain as the last of the seven ancient wonders of the world for 100,000 years to come.

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DETERMINATION OF CARBON DIOXIDE AND OTHER VOLATILES IN PYRITIC LIMESTONES BY LOSS ON IGNITION[1]

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ABSTRACT

A loss-on-ignition of limestones to 1000° C interrupted at 550° C for an intermediate weighing provided much of the information needed to calculate carbon dioxide gravimetrically but was complicated by the oxidation of pyrite, which is present in small quantities in many limestones. Upon oxidation pyrite formed Fe_2O_3 and oxides of sulfur which in turn reacted with $CaCO_3$ below the intermediate temperature to form $CaSO_4$ and cause a premature evolution of CO_2. It is necessary to determine sulfate sulfur and total sulfur in the raw sample and sulfate sulfur after loss-on-ignition at 1000° C. In limestones containing one percent or more pyrite, a sulfate determination after ignition at 550° C is necessary in addition to the other three sulfate determinations. These data enable calculation of a true CO_2 value and a single value for all other volatiles gravimetrically in each sample.

INTRODUCTION

Loss-on-ignition, with the collection of gases formed, has been used for many years to report volatile matter in mineral analyses (Hillebrand and Lundell, 1929). The determination of carbon dioxide in limestone samples by weight loss after the thermal decomposition of calcium carbonate has been carried out experimentally in the Geochemistry Laboratory of the Kansas Geological Survey since 1954. This method necessitates two weighings, one at an intermediate temperature before CO_2 begins to evolve from the decomposition of calcium carbonate and another at 1000° C in order to determine the weight after all CO_2 has been lost. The intermediate temperature and duration of ignition were determined from experimental evidence, and it was found that 550° C for 25 minutes volatilized most of the carbon, organic matter, and water, yet caused negligible loss of CO_2. Limestone samples that contain pyrite did not give a true value for CO_2, however, because error was introduced by the oxidation of pyrite to SO_3 and Fe_2O_3. The reaction of the SO_3 with $CaCO_3$ released CO_2 and formed $CaSO_4$, which is stable above 1000° C (Hillebrand, 1919; Hillebrand and Lundell, 1929; Kolthoff and Sandell, 1946). A series of experiments was initiated to determine a method of calculating a corrected CO_2 value and a single value for the other volatiles in pyritic limestones.

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EXPERIMENTAL METHODS

A Lindberg Model B-2 muffle furnace with an oxidizing plate and a chromel-alumel thermocouple controlled by a Lindberg Vacuum Tube Pyrometer Controller Model #291 was used for heating the samples. The control regulated the temperature within +20° C to −15° C of the selected temperature.

The samples used for this investigation were (1) limestones that contained microscopic disseminated pyrite crystals in amounts ranging from 0.3 to 1.2 percent, (2) National Bureau of Standards Argillaceous Limestone 1a (BS-1a), and (3) standards composed of Analytical Reagent $CaCO_3$ mixed by grinding with weighed portions of naturally occurring pyrite. All limestone samples and the sample of pyrite were ground to pass a 60-mesh sieve. The purity of the pyrite was 92.14 percent FeS_2.

The basic procedure consisted of placing a crucible containing a 2-gm sample dried at 105° C in the muffle at 550° C for 25 minutes; the sample was then removed and the first weight loss calculated. The sample was returned to the muffle, heated to a temperature of 1000° C, which was maintained for a minimum of one hour, and the percent loss of weight on heating from 550° C to 1000° C was calculated.

TABLE 1.—*Comparison of loss-on-ignition values at 550° C for 25 minutes with values at 600° C for 25 minutes*

Sample no.	58246* (Toronto)	58247* (Toronto)	BS-1a
% loss 550° for 25 min	1.42	1.29	1.03
% loss 1000° for 1 hr	30.87	38.38	33.47
% loss 600° for 25 min	2.02	1.87	1.62
% loss 1000° for 1 hr	30.28	37.76	32.81

* Limestone samples from a geologic formation in Kansas.

The effects of the following variables were studied: position in the muffle, type of crucible, and the temperature-time ratio for the intermediate weight. Placing the samples in various positions in the muffle did not affect the loss on ignition. Total loss-on-ignition values contrasting the weight loss in platinum and porcelain crucibles showed a maximum difference of 0.17 percent, indicating that either porcelain or platinum crucibles could be used. The maximum temperature-time ratio for the intermediate step was established to be 550° C for 25 minutes. Raising the temperature to 600° C for 25 minutes caused evolution of CO_2 (table 1). Heating at 550° C for 1.75 hours gave a loss on ignition of 1.46 percent (BS-1a). This indicates premature loss of CO_2 at the intermediate step. Therefore, the time the samples are to be held at 550° C cannot be increased beyond 25 minutes without some danger of losing CO_2 prematurely. Analytical Reagent grade calcium carbonate lost 0.3 to 0.4 percent in two hours, but only 0.1 percent in 45 minutes at 550° C.

An X-ray diffraction analysis on a sample of $CaCO_3$ mixed with pyrite ignited at 550° C indicated the presence of $CaSO_4$, $CaCO_3$, and Fe_2O_3. But for three limestone samples, sulfate analyses after the determination of loss on ignition at various temperatures (table 2) showed that although most of the oxidation of pyrite took place at 550° C (for 25 minutes), in samples containing more than one percent pyrite the reaction was not complete until 1000° C was attained.

It was thought that a lower intermediate temperature would allow an increase of time and thereby complete the oxidation of FeS_2 and the formation of $CaSO_4$ at the intermediate step; however, if the intermediate temperature were lower than 500° C, organic matter and water would not be completely evolved. Therefore, 520° C for 45 minutes was selected for the comparison analyses (table 3). No greater oxidation was obtained by this experiment.

Placing the dried samples in the cold muffle, raising the temperature to 550° C, and maintaining it for 25 minutes gave the same results as were obtained by placing the samples in the muffle at 550° C and maintaining the temperature for 25 minutes.

Evidence showed that in some limestones containing one percent or more pyrite the sulfide was not all oxidized and retained as $CaSO_4$, but a small portion evolved as SO_2 or SO_3. A standard composed of $CaCO_3$ and 3.12 percent pyrite showed after ignition at 550° C a sulfate value of 3.58 percent as SO_3. A total sulfur determination expressed as SO_3 on a sample of the standard of $CaCO_3$ and 3.12 percent pyrite ignited at 550° C showed a value of 3.45 percent. In these samples the theoretical value for total sulfur as SO_3 was 3.84 percent. Because the

TABLE 2.—*Sulfate values obtained after ignition at various temperatures*

Sample no.	58131* (Leavenworth)	598* (Merriam)	599* (Farley)
% SO_4 as SO_3 after 550° ignition	1.05	0.79	0.36
% SO_4 as SO_3 after 700° ignition	1.23	0.77	0.37
% SO_4 as SO_3 after 1000° ignition	1.60 (avg. 2)	0.82	0.37

* Naturally occurring pyritic limestones from three geologic formations in Kansas.

TABLE 3.—*Comparison of loss-on-ignition values at 550° C for 25 minutes with values at 520° C for 45 minutes*

Sample no.	58131* (Leavenworth)	598* (Merriam)	599* (Farley)
% loss @ 550° for 25 min	0.37 (avg. 2)	1.25 (avg. 2)	0.30 (avg. 2)
% loss @ 1000° for 1 hr	40.47	39.05	42.58
% loss @ 520° for 45 min	0.32 (avg. 2)	1.19 (avg. 2)	0.26 (avg. 2)
% loss @ 1000° for 1 hr	40.47	39.17	42.69

* Naturally occurring pyritic limestones from three geologic formations in Kansas.

sulfate value checked within reasonable limits with the total sulfur, which would include both sulfate and sulfide sulfur, any difference between the theoretical sulfate and actual sulfate must have been evolved as SO_2 or SO_3. It was found that for most limestones the amount of FeS_2 that was oxidized at temperatures above 550° C was negligible.

RESULTS AND CALCULATIONS

The stability of $CaSO_4$ at 1000° C permitted two loss-on-ignition weight determinations (550° C and 1000° C) on the same sample and the use of the fired sample for a sulfate determination. The sulfate after the 1000° C ignition, less the sulfate in the raw sample, gives the sulfate formed by the oxidation of pyrite. From these values it is possible to calculate a corrected 550° C loss of organic matter, and water, and a corrected CO_2 value.

The reaction involving sulfide retained as sulfate in the ignited sample proceeds according to the following formula:

$$4FeS_2+8CaCO_3+15\,O_2 \xrightarrow{550°\ C} 2Fe_2O_3+8CaSO_4+8CO_2$$

The 105° C to 550° C loss is incorrect by the difference in weight between O_2 gained and CO_2 lost, a net gain in weight. In limestone samples this gain is masked by the greater weight loss of water and other volatile material. The 550° C to 1000° C, or CO_2, loss is low because of the formation of stable $CaSO_4$ and loss of CO_2 below 550° C.

The CO_2 lost and replaced by SO_3 can be calculated by determining the moles of SO_3 formed and using the formula:

$$(b-a)\times \text{M.W. } CO_2=c$$

Where:

a = moles sulfate as SO_3 in the raw sample
b = moles sulfate as SO_3 after 1000° C ignition
$(b-a)$ = moles sulfate as SO_3 formed = moles CO_2 lost
c = weight CO_2 lost and replaced by SO_3

The total CO_2 value is the total of the 550° C to 1000° C loss plus c.

The calculation of 105° C to 550° C loss must take into consideration first the CO_2 lost below 550° C and expressed in the weight loss at 550° C, second, the oxygen gained during the reaction and retained in the sample as sulfate and ferric oxide, and third, any SO_2 or SO_3 lost as gas. The calculation of these three variables permits a report of a value for volatiles exclusive of sulfur and sulfur compounds. The CO_2 lost below 550° C has been calculated in equation 1. The weight gain of oxygen is calculated from the following relationship of the formula:

$$\frac{(b-a)15}{8} = \text{moles } O_2 \text{ gained}$$

Or

$$(b-a)\times 1.875\times 32=d$$

Where:

$(b-a)$ = moles CO_2 lost
d = wt. of O_2 gained at 550° C

Also from the formula:

$$(e-b)\times(4\times \text{M.W. S}-1\tfrac{1}{2}\times \text{M.W. } O_2)=f$$
$$(e-b)\times 20=f$$

Where:

e = moles of total SO_3 (sulfate and sulfide as SO_3) in the raw sample
b = moles sulfate as SO_3 after 1000° C ignition

f = wt. loss due to oxidation of FeS_2 to $Fe_2O_3+SO_2$

The corrected 550° C value = wt. loss at 550° C $+d-c-f$. When no SO_2 or SO_3 is evolved, e will equal b and f will be zero.

An optional method is available where necessary to determine more accurate values at 550° C and 1000° C on limestones containing one percent or more pyrite. Because the reaction is not complete in these limestones at 550° C, but continues beyond that temperature, it is necessary to determine loss on ignition on paired samples, extracting one at 550° C and one at 1000° C for sulfate determinations.

The following equations taken from the formula are used in the calculations:

$$(a_1-a)\times\text{M.W. } CO_2=C$$
$$(b-a_1)\times 1.875\times\text{M.W. } O_2=D$$
$$(a_1-a)\times 1.875\times\text{M.W. } O_2=D$$
$$(e-b)\times 20=f$$

Where:

a = moles sulfate as SO_3 in the raw sample
a_1 = moles sulfate as SO_3 at 550° C ignition
b = moles sulfate as SO_3 at 1000° C ignition
C = wt. CO_2 lost and replaced by SO_3 at 550° C
D = wt. O_2 gained at 1000° C
D_1 = wt. O_2 gained at 550° C
e = moles of total sulfate as SO_3 (sulfate and sulfide) in the raw sample
f = wt. lost in FeS_2 becoming $Fe_2O_3+SO_2$

The corrected CO_2 value = 1000° C loss $+C+D$.
The corrected 550° C value = 550° C loss $+D_1-C-f$.

DISCUSSION

The correction and method outlined above provide a rapid, accurate method for loss-on-ignition determination of CO_2 and volatiles in limestones containing 0.2 to 2.0 percent pyrite. Constant attention by the operator is not required, assuring a smooth flow of other determinations being carried out in the laboratory at the same time. The method further requires no special or additional equipment other than that found in most laboratories.

Two possible mechanisms for the reaction of FeS_2 and $CaCO_3$, when ground into an intimate mixture and heated, are that first the exothermic reaction of the oxidation of pyrite raises the temperature of the $CaCO_3$ in contact with it enough to trigger the endothermic loss of CO_2 with subsequent substitution of SO_3 to form a stable compound $CaSO_4$ and, second, that the oxides of sulfur formed on igniting pyritic limestones may react with small amounts of water that are retained in the limestone above 105° C forming H_2SO_4 when ignited at 550° C. The H_2SO_4 in turn would react with the $CaCO_3$ forming $CaSO_4$ and releasing CO_2 and H_2O. However, analytical reagent grade $CaCO_3$, dried at 550° C then mixed with 105° C dried pyrite, forms the $CaSO_4$ when ignited together at 550° C. Because the pyrite cannot be dried at 550° C without decomposition, it is possible a small quantity of water is retained in the pyrite and is forming the H_2SO_4.

The SO_2 loss from limestones containing more than one percent pyrite may result from the increased concentration of sulfur being oxidized. The resultant loss of CO_2 may cause a streaming effect which carries SO_2 from the upper part and exposed surface of the sample before it has time to react.

The application of the correction and method as defined is limited to limestones of relatively high purity and some dolomites containing only small amounts of clay minerals. Variation or adaptation of this method may prove applicable to impure carbonate rocks, but was not included in the scope of this study.

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EVALUATION OF HEAVY MINERAL SEPARATIONS USING ARTIFICIAL SAMPLES[1]

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ABSTRACT

In order to evaluate the method of heavy mineral separation used in Coastal Petroleum Company's laboratory work and to compare it with certain other gravity and centrifuge methods, 58 tests were conducted with artificial samples of known concentrations. Samples of 20 and 50 gm were compared in 1, 2, and 5 percent concentrations by the separatory funnel gravity method. Samples of 5-gm size were tested in the same concentrations using and comparing open funnel and separatory funnel gravity methods and double tube and separatory funnel centrifuge methods.

Relatively good results were obtained with both gravity methods and the separatory funnel centrifuge method since mean recovery was above 90 percent. It was found that for an artificial sample with a median diameter of 2.72 phi of the composition used and with the techniques followed the separatory funnel gravity method is adequate for our purpose, and it is as good as and perhaps superior to the centrifuge method.

INTRODUCTION

Coastal Petroleum Company has conducted a reconnaissance exploration for heavy minerals of some 400 miles of the west coast of Florida. In order to verify the results of this program, a test of the separation method was conducted using an artificial sample. A comparison was also conducted with fractions of the same sample on two gravity methods and two types of centrifuge separations. The present paper reports the results of these tests of laboratory methods using the artificial sample.

Equipment used in the study included a Sepor Micro-Splitter, U. S. Standard sieves, an O'Haus No. 2-034 triple beam stainless steel balance, a Mettler Type H-4 automatic balance, and an International Size 2, Model V centrifuge.

THE ARTIFICIAL SAMPLE

A quartz sand of low heavy mineral percentage was selected. It was separated three times by gravity in a separatory funnel to remove heavy minerals. It was then soaked in a weak hydrochloric acid solution to remove any shell or traces of phosphate.

Heavy minerals from a commercial operation in Florida were reseparated to insure purity. Ilmenite, rutile, zircon, monazite, and garnet were used.

The proportions of these minerals in the heavy fraction of the artificial sample were selected from the average percentages of 10 grain counts of samples taken during the field work offshore from Apalachicola, Florida. These average percentages were recalculated on a basis of 100 percent. The heavy fraction was assembled in three parts; one each for 1, 2, and 5 percent heavy minerals as shown by table 1.

In table 1 and subsequent tables the weights listed are those read directly from the scale of the automatic balance. They are given as read to provide as complete data as possible, and it is not pretended that they are a measure of the accuracy obtained in the test. The accuracy of the balance is 0.5 mg.

The heavy fractions were assembled with splits of the light fraction as shown in table 2.

A 20-gm split of the artificial sample was subjected to mechanical analysis in order to determine its gradation. The results are shown in table 3.

In table 3 the percentage figures given are direct readings from an automatic calculator and like the gram weights are not intended to reflect any degree of accuracy.

It was observed during the mechanical analysis that the majority of the heavy minerals were in the fractions retained on

[1] Publication authorized by the President, Coastal Petroleum Company. Manuscript received August 2, 1959.

TABLE 1.—*Weight composition of heavy fraction of artificial sample*

	Specific Gravity	Calculated Percent	1% H.M. gm	2% H.M. gm	5% H.M. gm
Ilmenite	4.5 –5	68.2	2.2506	4.5012	11.2530
Rutile	4.18–4.25	11.5	.3795	.7590	1.8975
Zircon	4.68–4.7	9.8	.3234	.6468	1.6170
Monazite	4.9 –5.3	9.1	.3003	.6006	1.5015
Garnet	3.15–4.3	1.4	.0462	.0924	.2310
		100.0	3.3000	6.6000	16.5000

the No. 170 and No. 230 sieves. Their grain size was therefore chiefly smaller than 0.125 mm and larger than 0.062 mm.

The cumulative percentages were plotted on probability paper, K & E No. 358-23. Sieve size was converted to phi units for this plot. The resulting curve fairly closely approximated a straight line. The 50 percentile was read as median, considering that the mean or average approximates the median. This value was 2.72 phi. Sorting was measured using the 16 and 84 percentile range which covers 68.27 percent of the spread of the cumulative curve and a standard deviation of two. Subsitutions in the standard deviation adopted by Inman (1952), Sigma $\phi = \frac{1}{2}(\phi 84 - \phi 16)$ gives a value for Sigma ϕ for this curve of 0.55.

Using portions or splits of the artificial sample, 58 tests were conducted, 24 to verify the method used in Coastal Petroleum Company's heavy mineral investigation and 34 in comparison of gravity and centrifuge separation methods.

TESTS OF COASTAL'S GRAVITY SEPARATIONS

The majority of the samples obtained with the underway sampler were about 20 gm. It was desired to determine if similar results would be obtained with larger samples such as 50 gm. These two weights were selected and tests run of the 1, 2, and 5 percent fractions of the artificial sample. Four samples of each were run and the results averaged. Separation was in 250-ml separatory funnels. This type of funnel has been described by Clemmons, Stacy, and Browning (1957).

The method used is briefly outlined as follows:

1. Separatory funnels in a rack were filled ¾ full with tetrabromoethane and the whole sample added.
2. The sample was stirred thoroughly with glass rods. The rods were carefully washed with the heavy liquid.
3. Allowed to stand 15 minutes or longer if not well settled.
4. Stirred well a second time and allowed to stand another 15 minutes.
5. Stirred a third time, and allowed to stand another 15 minutes or until well settled.
6. Heavies drawn off through petcock onto filter paper in open funnels.
7. A little more tetrabromoethane added

TABLE 2.—*Assembly of the artificial sample*

	1% Heavy Minerals	2% Heavy Minerals	5% Heavy Minerals
Heavy Fraction	3.3 gm	6.6 gm	16.5 gm
Light Fraction	326.7 gm	323.4 gm	313.5 gm
Whole sample	330.0 gm	330.0 gm	330.0 gm

TABLE 3.—*Grain size analysis of the artificial sample*

U. S. Standard Sieve No.	Opening mm	Weight Retained (gm)	Percent Retained	Cumulative Percent Retained
25	0.71	.008	.0400	.0400
35	0.50	.058	.2900	.3300
45	0.35	.235	1.1751	1.5051
60	0.25	1.958	9.7909	11.2960
80	0.177	5.652	28.2628	39.5588
120	0.125	6.648	33.2430	72.8018
170	0.088	4.543	22.7173	95.5191
230	0.062	.785	3.9254	99.4445
Pan		.111	.5550	99.9995
		19.998	99.9995	

washing down sides of separatory funnel; sample stirred again and allowed to settle.

8. Heavies drawn off again onto filter paper.
9. Sample washed 3 times with alcohol or benzene.
10. Heavies dried in oven.
11. Heavies weighed and percentage calculated.
12. Lights also let down on filter paper and washed, dried, and saved for reference.

Twenty-four separations were conducted. The separation time for a rack of six samples was approximately one hour. The results of this work are shown by table 4.

COMPARISON OF CENTRIFUGE AND GRAVITY METHODS

Centrifuge Methods

A centrifuge was made available at the

TABLE 4.—*Data obtained by gravity separatory funnel separation of 20 and 50 gm artificial samples with 1, 2, and 5 percent heavies*

20-gm Samples				50-gm Samples			
1% Heavies—Prepared Concentrations							
Sample No.	Weight of Heavies Recovered			Sample No.	Weight of Heavies Recovered		
	Gm	Percent	% of Total (1%)		Gm	Percent	% of Total (1%)
1a	0.198	0.990	99.00	1a	0.495	0.990	99.00
b	0.196	0.980	98.00	b	0.497	0.994	99.40
c	0.191	0.955	95.50	c	0.501	1.002	100.20
d	0.191	0.995	95.50	d	0.502	1.004	100.40
Mean		0.970	97.00	Mean		0.998	99.75
2% Heavies—Prepared Concentrations							
Sample No.	Weight of Heavies Recovered			Sample No.	Weight of Heavies Recovered		
	Gm	Percent	% of Total (2%)		Gm	Percent	% of Total (2%)
2a	0.395	1.975	98.75	2a	1.005	2.010	100.50
b	0.398	1.990	99.50	b	0.985	1.970	98.50
c	0.404	2.020	101.00	c	1.004	2.008	100.40
d	0.393	1.965	98.25	d	1.002	2.004	100.20
Mean		1.990	99.38	Mean		1.998	99.90
5% Heavies—Prepared Concentrations							
Sample No.	Weight of Heavies Recovered			Sample No.	Weight of Heavies Recovered		
	Gm	Percent	% of Total (5%)		Gm	Percent	% of Total (5%)
5a	1.004	5.020	100.40	5a	2.504	5.008	100.16
b	0.992	4.960	99.20	b	2.505	5.010	100.20
c	0.995	4.975	99.50	c	2.505	5.010	100.20
d	0.992	4.960	99.20	d	2.495	4.990	99.80
Mean		4.988	99.75	Mean		5.002	100.09
Mean Recovery 1, 2, and 5% samples			98.65	Mean Recovery 1, 2, and 5% samples			99.91

Geology Department of Florida State University. Qualitative studies of heavy minerals have been in progress there in which clear splits are obtained for microscopic study of heavy minerals. Quantitative results, when obtained, are derived from procedures other than weight percentages.

An attempt was made to adapt this method to weight percentage quantitative analysis. Two glass tubes were fitted into a trunnion cup of the centrifuge. The outer one was similar to a test tube and was cushioned at the bottom in the trunnion. The inner tube was smaller, shorter, tapered, and open at the lower end. Its upper end was held in position by a collar which fitted over the upper end of the outer tube. This arrangement was probably very similar to that described by Twenhofel and Tyler (1941) as the Wolff method.

Five-gm samples, similar in size to those used by Rittenhouse and Bertholf (1942) were used. Again 1, 2, and 5 percent samples were employed, a total of 16 samples. The technique used was as follows:

1. The outer glass tube was filled ¾ full with tetrabromoethane and a 5-gm sample placed in the inner tube.
2. The sample was agitated gently. Too vigorous stirring was found to send the lights down to the open bottom of the inner tube and up on the outside in the outer tube, causing difficulty in later removing the heavies.
3. Sample placed in centrifuge which was then closed and started. Five minutes were allowed for the machine to gather speed to 2000 rpm, and 10 minutes were allowed at that speed.
4. Machine stopped and heavies removed from outer tube, washed with alcohol or benzene, dried, and weighed as in other method.

These samples were centrifuged in pairs. Total separation time was about 20 minutes. Positive separations between the lights and heavies were experienced. However, because many heavies remained in the lights, poor recovery of known percentages resulted. Recovery was roughly 30 to 80 percent. Four of the 16 samples were rerun in the centrifuge. Recovery was improved somewhat, but it did not exceed 90 percent.

From observation it appeared that the 5 percent samples had poorer recovery than the 1 and 2 percent samples. This point was not pursued further, however, and additional testing is suggested.

It is possible also that the 5-gm samples were too large for the tubes used. A 2- or 3-gm sample may have produced better results. Another possibility is that better results might have been obtained if the centrifuge had been stopped during the 10 minutes and stirred.

With the apparent failure of the double tube method for our purpose, another technique was sought. A search of the literature indicated several techniques such as those described by Taylor (1933), Krumbein and Pettijohn (1938), and Bertholf (1940). However, our tests were limited to techniques possible with glassware available with the centrifuge at Florida State University.

A 125-ml separatory funnel was selected. The technique was as follows:

1. 25 ml of tetrabromoethane were poured into the 125-ml funnel and a 5-gm sample added.
2. The mixture was well stirred with a glass rod and the stirring rod was washed down with about 3 ml tetrabromoethane.
3. The separatory funnel was placed snugly in the trunnion carrier. The petcock was checked to be sure it was tightly closed. The petcock was placed upward in the centrifuge.
4. The timer of the centrifuge was set for approximately 12 minutes. The centrifuge was allowed to gain speed slowly to 900 rpm. This speed was maintained for 10 minutes. Then the centrifuge was allowed to come to a stop naturally without braking.
5. The heavies were removed from the separatory funnel on filter paper, washed, dried, and weighed.

Six samples were run in pairs. Total separation time for a pair was about 15 minutes. It was found that if too much heavy liquid was used in this separatory funnel the weight of the liquid under centrifugal force loosened the petcock and the liquid was lost. This funnel was not designed to be used at a high speed. The recommended high is 1200 rpm. It was believed that breakage and loss of heavy liquid might occur at even this speed and this speed was not attempted.

Gravity Methods

For comparison, similar 5-gm samples were tested in the same separatory funnels by gravity. The technique was essentially the same as the one previously described for this method except that the time was reduced slightly because of the smaller size of the samples. Stirring was at 10 minute intervals. Six samples were separated in pairs. The separation time for a pair was 35 to 40 minutes.

Further comparison was provided by testing similar 5-gm samples in an open funnel. The type of funnel was similar to those shown by Milner (1952). The technique was the same as for the 5-gm separatory funnels. Six separations were run in pairs. The separations were conducted with the 5-gm separatory funnel gravity samples during the same time interval of 35 to 40 minutes.

Comparison

The results of this three way comparison are shown by table 5.

RESULTS

In testing the method used in Coastal Petroleum Company's heavy mineral field program (gravity separatory funnel separations of 20-gm samples), the recovery of heavies of the artificial sample (table 4) was found from the means to be from 97 to 99 percent. Recovery was over 99 percent in samples of 2 percent and 5 percent concentration. Percent recovery increased with increased heavy mineral concentration. The average recovery of the three concentrations was 98.6 percent.

The 50-gm samples (table 4) showed recovery of more than 99.9 percent. This seems to indicate that recovery was more complete in 50- rather than 20-gm samples. The differences are little more, however, than the expected inaccuracy of the tests. The mean of the concentrations of the 50-gm samples varied from 99.8 to 100 percent and percent recovery appeared to increase with increased concentration.

The 5-gm samples had means for the three methods (table 5) of 94.8, 97.9, and 96.1 percent, averaging 97.3 percent. This was a little less than the 20- and 50-gm samples, again indicating better recovery with the larger samples.

TABLE 5.—*Data obtained by comparison of open funnel gravity, separatory funnel gravity, and separatory funnel centrifuge methods of separation of 1, 2, and 5 percent artificial samples*

Concentration	Open Funnel Gravity Method					Separatory Funnel Gravity Method					Separatory Funnel Centrifuge Method				
	Sample No.	Weight Gm	Weight Recovered			Sample No.	Weight Gm	Weight Recovered			Sample No.	Weight Gm	Weight Recovered		
			Gm	%	% of Total			Gm	%	% of Total			Gm	%	% of Total
1 Percent Heavy Minerals	1a	5.156	0.047	0.912	91.20	2a	4.910	0.048	0.978	97.80	3a	4.710	0.050	1.062	106.20
	b	5.050	0.047	0.931	93.07	b	5.145	0.050	0.972	97.20	b	4.870	0.042	0.862	86.20
	Mean			0.922	92.14	Mean			0.975	97.50	Mean			0.962	96.20
2 Percent Heavy Minerals	4a	4.912	0.100	2.036	101.86	5a	5.032	0.100	1.987	99.35	6a	4.910	0.098	1.996	99.80
	b	4.940	0.088	1.781	89.05	b	5.178	0.098	1.893	94.65	b	5.035	0.094	1.867	93.35
	Mean			1.908	95.42	Mean			1.940	97.00	Mean			1.932	96.58
5 Percent Heavy Minerals	7a	4.730	0.234	4.947	98.94	8a	5.005	0.249	4.975	99.50	9a	5.052	0.245	4.850	97.00
	b	4.864	0.230	4.729	94.58	b	5.020	0.248	4.940	98.80	b	5.010	0.235	4.691	93.82
	Mean			4.838	96.76	Mean			4.958	99.15	Mean			4.770	95.41
Means of 1, 2 and 5 Percent Samples					94.77					97.88					96.06

The results of the three method comparison of 5-gm samples (table 5) were also of interest. As expected, the open funnel gravity method gave consistently lower results. The means varied from 92 to almost 97 percent, with an average of almost 95 percent. Percent recovery increased with concentration. The lowest open funnel recovery recorded was 89 percent (sample 4b).

The range of the means of the separatory funnel gravity method was from 97 to 99 percent. The 2 percent sample, rather than the 1 percent, was the low mean. The 5 percent mean was again the highest. The overall average here was the highest for the three methods—almost 98 percent.

The range of the means of the separatory funnel centrifuge method was from more than 95 to more than 96 percent with an average of 96 percent. The increase with concentration holds with the 2 percent over the 1 percent, but the 5 percent mean was lower.

CONCLUSIONS

Relatively good results were obtained with three methods of separation as the means all exceeded 90 percent recovery.

The separatory funnel gravity method was indicated by this work to be the highest in recovery of the heavy minerals from an artificial sample of the composition used for the techniques followed. It is believed that this method is adequate for the purpose for which it was used in our work. It is also well adapted to samples of 50-gm size and with a little less recovery to a 5-gm sample.

The centrifuge method ran a close second. With certain other equipment not here available or with techniques not here performed, it is probably just as good or better.

Percent recovery usually, but not always, increased slightly with heavy mineral concentration. There was also an increase on the basis of sample size; the larger the sample, the more complete the percent recovery.

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NOTES

THE TERM GRAYWACKE[1]

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It may be regarded as a sign of the times that no less than four papers were presented at the Annual Meeting of the Geological Society of America recently, deploring the careless writing of many geological papers and particularly the misuse of words. Similar complaints about faulty diction and loose terminology have been heard not infrequently in Britain for more than a decade. Many examples of the confusion that has resulted could be cited. During the discussions at St. Louis in 1958, referred to above, the term *Graywacke* was singled out as an example of bad practice. Since the muddle that has again developed over the application of this term is not creditable to those of us who are especially interested in the study of sedimentary rocks, I venture to suggest that the word should either be abandoned or that its use should be restricted to a precise connotation. Some discussion may be necessary to achieve this, but it is a matter of sufficient importance and urgency to bring forward, even if it has to be by "unofficial" conference, at the forthcoming International Geological Congress at Copenhagen.

At the end of last century the word graywacke (greywacke, Grauwacke) had fallen almost entirely into disuse. However, when a Committee on Sedimentation was appointed by the National Research Council of the U.S.A. after World War I the name came forward for consideration. Lengthy international discussion by correspondence showed that many geologists were in favour of discontinuing its use on the ground that so many different meanings had been attached to it. Nevertheless, some petrologists advocated its retention because there appeared to be no other term available for the group of rather coarse-grained, dark, polygenic rocks to which it had been applied in the early days. General agreement on a definition was secured, but only with difficulty, by the indefatigable chairman of the sub-committee on medium-grained rocks, Victor T. Allen (see final Report, 1936, p. 28, 46). He remarked that " . . . the history of the term graywacke is one of periodic revival after years of disuse. This seems to indicate that geologists will continue to employ the term and additional meanings will be added to the confusion that surrounds it." He also expressed his opinion that the definition suggested would "not meet the approval of everyone."

Ancient British geologists had brought from Germany the term *Grauwacke* which became Anglicized to graywacke or greywacke (and occasionally during the past few decades acquired a Gallic e acute) in order to give a name to certain British Paleozoic rocks. These were of mixed grade (that is, were ill-sorted) and contained a considerable variety of constituents; in many instances incidentally, the rocks bore no close resemblance to the original German grauwackes. The term was used to denote formations, and thus served to meet a stratigraphical rather than a petrographical need. Nonetheless, from the time Wm. Humble gave it a place in his geological dictionary of 1840 it was used from time to time throughout the nineteenth century as a rock term (for example, by Lyell, Geo. Roberts, Page and others), but with decreasing enthusiasm as the microscope gave more precision to petrology.

The definition eventually proposed by the American Committee was as follows:

A sandstone composed of 33 or more percent of easily destroyed minerals and rock fragments derived by rapid disintegration of basic igneous rocks, slates and dark colored rocks. It may or

[1] Manuscript received November 16, 1959.

may not be intensely indurated or metamorphosed.

This definition owed much to G. W. Tyrrell's writings and advice, but was not drafted by him. Also, it is not repugnant to J. S. Flett's definition in the Encyclopedia Britannica,[2] or to those of A. Harker and others in their widely-used textbooks. Although I am on record in the Report as being among those who at first favoured the abandonment of the term, I should say that eventually I accepted the definition—with some misgivings! It may now be felt, as it was then, that the definition states too little about the qualities of the rock. In any event, the confusion that we feared has increased steadily since 1936.

In 1958 F. B. Van Houten did a valuable service by translating for the Geological Society of America Reinhard Helmbold's account of the petrography of the graywacke from Tanner in the Harz mountains, where it was first recorded. Three varieties, coarse, medium, and fine, are described, and the chemical analyses are, it so happens, closely similar to Tyrrell's average analysis for graywackes. The sorting of the grains is poor, a great variety of minerals and rock fragments is recorded, and a substantial quantity of micaceous and chloritic matrix is in evidence. The irregularity in the shape and size of the grains is emphasized. Chemical analyses show that soda is in excess of potash, and ferric iron exceeds ferrous iron, but magnesia does *not* exceed lime. Thus it would seem that here we have a rock group which should properly bear a distinctive name.

The most serious aspect of the revival of the term graywacke during the past twenty years is that few geologists agree with each others' definitions, and still fewer remember to state, for the benefit of the reader, the particular connotation they give to the word. Some examples of modern usage are added below in the hope that their citation may assist in obtaining prompt attention to the problem.

It is often declared that graywackes are essentially coarse-grained rocks, i.e. conglomerates, breccias, and sandstones (2 mm to 0.06 mm e.s.d.); withal, they must be poorly sorted sediments, that is contain a good proportion of many grades, including silt and clay. Unless the discussion is specifically petrographical, little more is usually said about the mineralogy of the sand and smaller grades than that quartz, feldspar, mica, and chlorite are present. As an example of the difficulties that arise from inadequate definition, I may refer to a recent paper on the Aberystwyth Grit Series (Lower Silurian) in Wales, valuable for its discussion of textural features and surface markings. The Series consists of alternating beds of graywacke and mudstone (or so-called shale). The graywackes are defined only by their grading, graphs being supplied in illustration. In three examples recorded as characteristic because of the poor sorting, one contains 70 percent silt and clay, the second 62 percent, and the third 87 percent (of which a considerable proportion is clay grade). From their mechanical composition, therefore, these rocks are by usual acceptance actually siltstones or mudstones. Obviously, the naming does not fit the analyses. Two other graphs of graywackes that are added later represent well-sorted fine- to medium-grained sandstones.

Of another well-known series of quartzose or feldspathic sandstones of Middle Silurian age in north and central Wales, the author states that they "clearly belong" to the graywacke facies and consist of "a series of greywackes, siltstones, and mudstones," but the term greywacke is not defined otherwise than by implication. However, it is clear that the author regards poor sorting, angular to sub-rounded shape of grains, variety of rock-fragments and minerals, and the presence of curious surface markings (flutings, grooves, casts, etc.) as criteria. Graded and "ungraded" bedding and current-bedding are all said to be common, and a shelly shelf fauna is recorded. Many other writers also consider the presence of surface markings of the type mentioned to be characteristic; some think that graded bedding is essential and that current-bedding is absent in true graywackes, but some dispute this.

Graywackes are frequently stated to be

[2] In the 13th edition. The note in the 14th edition (1929) is over the initials P.G.H.B.

deep water deposits. One author at least has declared them to be the product of re-sedimentation of unconsolidated deposits beyond the reach of sand-pushing currents, but what features were to be regarded as diagnostic is not clear. In dealing with graywackes in one of the popular Penguin handbooks, prepared for general readers as well as scientists, the author states that "practically all fossil remains of sea-floor organisms are missing, and none of the normal inhabitants of the shallow-seas are found . . . the only organisms found in beds of greywackes are those where body-form is appropriate to a floating existence, and which one might expect to find in the deposits of an ocean deep." Evidence of breaks in sedimentation and of surfaces of erosion are, according to some geologists, absent from true graywackes.

It will not have escaped notice from these observations that there is a growing tendency for accessory features and inferences about the mode and place of origin to become diagnostic instead of supplementary to the petrographical criteria formerly regarded as essential.

Now we find that the wheel has almost turned full circle. The Cambrian sandstones (Hell's Mouth Grits) in the country west of Snowdonia have recently been re-described thus: "The Grit beds exhibit the characteristic textures and structures of greywackes, but differ from the normal type in being relatively well sorted and commonly laminated. They are dominantly lithic (low-rank) greywackes with a small percentage of feldspathic greywackes." It will thus be seen that modern fashion requires that the rocks are still to be termed graywacke but are styled abnormal because they resemble ordinary sandstone!

Already there are signs that the connection between geosynclines and graywackes is being increasingly emphasized, notwithstanding the obvious danger of reasoning in a circle. Thus, once the existence of a geosyncline is suspected, efforts are made to strengthen the case by the discovery of graywackes: also, the presence of graywackes leads to the search for the geosyncline, the early stages of upheaval of which gave rise to them. Incidentally, references that are often made abroad to the outstanding example of a geosyncline shown by the characters of the Older Paleozic rocks of Wales (which land was part of the adopted home of the *Grauwacke* a century ago) probably result from the enthusiastic advocacy of a distinguished Welsh geologist. On the highly debatable question of its reality, opinions differ in Britain and elsewhere. A well-known Netherlands geologist regards it as the prototype of a new kind of geosyncline characterized by its symmetrical form, shelf-margins, and the quantity of graywackes in its central part. But opinions also differ, as noted above, about the supposed graywackes! This is not the place to enter into details of the argument whether or not still another, hitherto useful, term is being misapplied; in any event it does not help us to reach agreement about the identity of a graywacke when we meet it.

To summarize:

1. The term was, and still presumably is, intended for use in the field. If we insist that a certain range of chemical composition is to be a criterion, we must sacrifice the idea of naming the rock in the field (although we may make conjectures about the proportions of some elements from the macroscopical appearance).

2. The opinions recorded in the foregoing indicate that the mechanical composition (or sizing), the mineralogical composition (including that of the rock-fragments), the shape of the grains, and certain structural and textural features that are considered to throw light on the mode of formation and source are criteria that may be adopted by some authors but waived by others. In many cases opinions are flatly contradictory.

3. The most favoured criteria appear to be (a) sizing, from conglomerates and breccias to sandstones; (b) poor sorting of grains, with a substantial quantity of fine matrix; (c) rock and mineral fragments, remarkably variable; (d) shape of grains, predominantly angular to subangular.

Rocks possessing the characters stated in the last paragraph are widely distributed throughout the world, and are worthy of a distinctive name. If the term graywacke is considered to be now too vague, the need

must be met by introducing a new one. Many such can doubtless be devised. *Usamerite*, for example, carries an implied compliment to the country where the study of sedimentary rocks has been so actively encouraged; it has the advantage of a form which may assist (by its alternation of vowels and consonants) experimental computation when that comes, in due course, to be developed.

NOTES ON CLASSIFICATION OF CARBONATE ROCKS ON BASIS OF CHEMICAL COMPOSITION[1]

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The first attempt to classify carbonate rocks on the basis of CaO/MgO ratio was apparently made by S. G. Vishnyakov (1933, in Frolova, 1959, p. 34). His classification is presented in table 1.

Teodorovich proposed a similar classification in 1935 (Teodorovich 1950, p. 126), which is shown in table 2.

Of the proposed classifications in the United States, Pettijohn's (1949) chemical scheme allowed an accurate classification of the intergradations of carbonate rocks. In 1955, Guerrero and Kenner (p. 46) proposed a

[1] Manuscript received August 28, 1959.

TABLE 2.—*Teodorovich classification of calcite-dolomite series*

Type	Calcite content, %	Dolomite content, %
Limestone	100–95	0–5
Slightly dolomitic limestone	95–80	5–20
Medium dolomitic limestone	80–65	20–35
Highly dolomitic limestone	65–50	35–50
Highly calcitic dolomite	50–35	50–65
Medium calcitic dolomite	35–20	65–80
Slightly calcitic dolomite	20–5	80–95
Dolomite	5–0	95–100

TABLE 1.—*Vishnyakov's classification of limestones and dolomites*

Type	$CaCO_3$ content, %	$CaMg(CO_3)_2$ content, %	CaO/MgO ratio
Limestone	95–100	5– 0	50.1–∼
Slightly dolomitic limestone	75– 95	25– 5	9.1–50.1
Dolomitic limestone	50– 75	50–25	4.0– 9.1
Calcitic dolomite	25– 50	75–50	2.2– 4.0
Slightly calcitic dolomite	5– 25	95–75	1.5– 2.2
Dolomite	0– 5	100–95	1.4– 1.5

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CHILINGAR, G. V., 1957, Classification of limestones and dolomites on basis of Ca/Mg ratio: Jour. Sedimentary Petrology, v. 27, p. 187–189.

FROLOVA, E. K., 1959, On classification of carbonate rocks of limestone-dolomite-magnesite series: Novosti Neft. Tekhniki (Geology), no. 3, p. 34–35. (This periodical is edited by the foremost Soviet petroleum geologist I. O. Brod, and is available from the writer.)

GUERRERO, R. G., AND KENNER, C. T., 1955, Classification of Permian rocks of Western Texas by a versenate method of chemical analysis: Jour. Sedimentary Petrology, v. 25, p. 45–50.

PETTIJOHN, F. J., 1949, Sedimentary rocks. Harper and Brothers, New York, 526 pp.

TEODOROVICH, G. I., 1950, Lithology of Paleozoic carbonate rocks of Ural-Volga region. Izd. Akad. Nauk SSSR, Moscow-Leningrad, 213 p. (Available from George V. Chilingar.)

TABLE 3.—*Frolova's classification of calcite-dolomite-magnesite series*

Type	Calcite content, %	Dolomite content, %	Magnesite content, %	CaO/MgO ratio
Limestone	95–100	5– 0	—	50.1 –∼
Slightly dolomitic limestone	75– 95	25– 5	—	9.1 –50.1
Dolomitic limestone	50– 75	50–25	—	4.0 – 9.1
Calcitic dolomite	25– 50	75–50	—	2.2 – 4.0
Slightly calcitic dolomite	5– 25	95–75	—	1.5 – 2.2
Dolomite	0– 5	100–95	—	1.4 – 1.5
Very slightly magnesian dolomite	—	100–95	0– 5	1.25– 1.4
Slightly magnesian dolomite	—	95–75	5– 25	0.80– 1.25
Magnesian dolomite	—	75–50	25– 50	0.44– 0.80
Dolomitic magnesite	—	50–25	50– 75	0.18– 0.44
Slightly dolomitic magnesite	—	25– 5	75– 95	0.03– 0.18
Magnesite	—	5– 0	95–100	0.00– 0.03

modification of Pettijohn's quantitative scheme of calcite-dolomite-clay series. Guerrero and Kenner (1955, p. 48) also proposed a classification of limestone-dolomite series on the basis of CaO/MgO molar ratio.

Inasmuch as many carbonate rocks contain magnesite in addition to calcite and dolomite, a more elaborate classification of calcite-dolomite-magnesite series was recently proposed by Frolova (1959, p. 35) and is presented in table 3.

The writer (Chilingar, 1957, p. 187) also proposed a classification of limestones and dolomites on the basis of Ca/Mg ratio, with some divisions dictated by the mode of formation. Thus this classification is of value because it is a combination of genetic and compositional schemes.

REVIEWS

The Study of Rocks in Thin Section, by W. W. Moorhouse, 1959. Pp. xvii+514, 226 figures, 18 plates, 17 tables, $6\frac{1}{4}\times9\frac{1}{4}$ in., cloth, Harper and Brothers, Geoscience series, New York. Price $8.00.

As the author himself points out, his book is essentially a petrography laboratory manual. Even with such a restriction, the title is somewhat misleading because it should read: "The study of igneous and metamorphic rocks in thin section with an appendix on sedimentary rocks." There appears to be a tradition that half of the content of this type of textbook must be devoted to igneous rocks and that metamorphic and sedimentary rocks share the remaining space unequally. Such an approach is, in the reviewer's opinion, no longer warranted, from a pure scientific point of view or from an economic one, as shown, for instance, by the great development of research in carbonates and in clays.

The book, in spite of its basic lack of balance, has outstanding qualities among which is the convenient association in one volume of a summary of the methods of optical mineralogy and descriptions of the most frequent rock-forming minerals with identification tables and descriptions of the common rocks.

The classification and discussion of the igneous rocks is clear and practical and the author has displayed a rare ability in such an approach which is a clear expression of his professional experience. It is appropriate here to stress the excellent quality of the illustrations, both line-drawings and photomicrographs, and particularly the idea of illustrating most of Ward's *100 American Rocks*. A very welcome innovation is the attention payed to altered rocks since they represent a real problem for the field geologist.

The sedimentary rocks have been discussed as if they were a burden to be disposed of as quickly as possible. Because the economic standpoint is somewhat stressed in the book, it is surprising to find, for in-

stance, the treatment of the evaporites and of the phosphates reduced to about three pages. As a result of such an abbreviated analysis of the sedimentary rocks, many statements have been grossly oversimplified and several rock-types have definitions which are shorter than those found in common dictionaries and appear certainly unsufficient even for a laboratory manual.

The discussion of metamorphic rocks is another interesting and excellent part of the book and appears once more to reflect the personal experience of the author. The chapters on dynamic, thermal and regional metamorphisms are new and certainly represent a rather unique and original contribution; of the same quality is the discussion of metasomatism and petrography of ores.

Professor Moorhouse's book has outstanding qualities which largely predominate over its shortcomings; it will be appreciated by students and field geologists. However, in the opinion of the reviewer, such a book seems to point out, once more, a critical problem in the ever increasing complexity of petrography, mainly the necessity of team work for an adequate coverage of all fields of petrography.

ALBERT V. CAROZZI
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The Lognormal Distribution, with special references to its uses in economics, by John Aitchison and J. A. C. Brown, 1957. Pp. xv+176; 33 figs., 44 tables, 6×9.7 in.; cloth. Cambridge University Press, Cambridge. Price $6.50.

This special-purpose statistics book, despite the statement about "economics" in the sub-title, should be of very great interest to the student of sediments. In fact, much of the history covered in the early part of the book has to do with studies in sedimentology and the references to familiar names—such as Krumbein—are numerous.

The lognormal distribution is defined by the authors as the distribution of a variate whose logarithm obeys the normal law of probability. This can be illustrated in sedimentology by Krumbein's phi-scale, which allows lognormal size distributions to be handled as if they were normal (that is, Gaussian). It should not be assumed, however, that the concepts of the normal distribution and the notion of logarithms completely "wrap up" the lognormal distribution. The utility of the phi-scale does not prove an identity between the two types of distribution. In fact, the lognormal distribution has certain properties which can not be derived from either the normal distribution or the log tranformation. Among these are additional parameters and some special difficulties in estimation procedures.

The authors review the "theory of breakage" as adequate to account for the lognormal distribution and give examples of how this theory can be applied. Much of the discussion has to do with granular matter (that is, sediments). They also review various elementary lognormal distributions, including the following: (a) two parameter distribution, defined by the mean and the standard deviation; (b) three parameter distribution, which also requires the concept of the "threshold" as a lower boundary; and (c) four parameter distribution, including the notion of an upper boundary. The four-parameter distribution proves particularly intractable.

One chapter is devoted to artificial lognormal samples and tests of lognormality. The latter includes a discussion of graphic tests (the quick-and-easy method of plotting the cumulative frequency curve on probability paper) which are useful but which do not permit a numerical assessment of reliability.

Truncated and censored distributions and the special problem of zero observations are treated in one chapter. This is perhaps not so important in most routine sediment studies where mixing of materials does not generally imply the suppression of either one or the other. In some types of earth science study, however, it *is* important; for example, when geomorphic surfaces are mixed in the same area, the addition of one generally implies the partial destruction of another. Truncated distributions may appear, for example, where arbitrary limits are imposed, such as in pebble diameter sampling.

Of the 13 chapters in the book, nine are more-or-less of general geological interest (such as the examples cited above), one has to do with computer problems (specifically

the EDSAC), and two are more nearly limited to economics (probit analysis; distribution of incomes).

The reader will never doubt that this is a statistics book, written primarily for students of statistics; for complete understanding, it requires mathematics on the level of partial differentials. It should be "must" reading, however, for any geologist who wishes to clarify his work by means of numerical evaluations.

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An Introduction to Fluvial Hydraulics, by Serge Leliavsky, 1955, corrected impression, 1959. Pp. xii+257, 88 figs., list of symbols, name index, and subject index; 5.5×8.75 in., cloth. Constable and Co., London. Price $7.50.

In the author's words (author's preface), "this book is an attempt at a short but comprehensive survey of the various theories, methods and facts bearing upon the flow of water in erodible channels, which are themselves the natural creations of this flow." In the light of this goal, the book must be considered an unqualified success. It brings together, in a single, neat, convenient package, practically all of the pertinent material on the subject. And it does so with clarity, force, logic, and a charming manner with the English language.

Dr. Leliavsky has been, at one time or another, director of the designing service, reservoirs and Nile barrages department, Ministry of Public Works, Cairo; chief of designing office, irrigation projects department, Ministry of Public Works, Cairo; professor of irrigation design, Royal School of Engineering, Guiza, Cairo; head of the bridges department, Egyptian State Railways, Cairo; and head of the hydraulic section of lock and dam construction works on the Dnieper Falls, Russia. His familiarity with the literature produced by all nations, including Russia, is obvious. And he has clearly drawn inspiration for his work from Nicolas de Leliavsky.

The ten chapters deal with the following subjects: (1) theoretical versus empirical approach; (2) difference between traction and suspension; (3) correlation between surface slope and particle size; (4) dunes and ripples; (5) pick-up velocity, drag and lift, as criteria of scour; (6) bed-load in the light of the drag theory; (7) side-slope stability in the light of the drag theory; (8) the non-parallelism principle in the interpretation of three-dimensional water flow in rivers; (9) sediment suspension explained mechanically; and (10) the empirical approach to the sediment transportation problem.

In a brief review it would be impossible to give a complete and clear picture of the book. Too many subjects are handled (and handled well). In only a few places might one quibble over definitions or usages. One of these occurs on page 3, where "saltation" is listed as though it were a prominent method of transportation; another on page 5, where the author "suggests" that steep slopes are symptomatic of higher velocities.

The terminology is not, at all places, standard sedimentological or geomorphological jargon; it does not appear, however, that Leliavsky has lost anything in clarity by using plain English.

The book does not include certain pertinent references of the last few years. Examples of omissions are the classic discussion of the graded stream by Mackin (1948) and the study of various stream parameters by Leopold and Miller (1956). On the other hand, it is more comprehensive and more useful than a similar (and excellent) treatment by Blench (1957).

No student of clastic sediments or river morphology can afford to pass Leliavsky's book by.

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BLENCH, T., 1957, Regime behaviour of canals and rivers. Butterworths Scientific Publications, London; 138 p.

LEOPOLD. L. B., AND MILLER, J. P., 1956, Ephemeral streams—hydraulic factors and their relation to the drainage net, U. S. Geological Survey Professional Paper 282-A, 35 p.

MACKIN, J. H., 1948, Concept of the graded river: Geol. Soc. America Bull., v. 59, p. 463–512.

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The Earth Beneath the Sea, by Francis P. Shepard, 1959. Pp. xii+275, 113 figs., 5½×8¼ in., cloth. The Johns Hopkins Press, Baltimore, Maryland. Price $5.00.

Almost ¾ of the surface of the earth lies beneath the seas. In the 10-year period following the publication of Dr. Shepard's book *Submarine Geology*, in 1948 and the writing of *The Earth Beneath the Sea*, there has been an increased interest in the seas and in the earth beneath the seas. This interest has not been confined to geologists and other scientists but is shared by countless persons who have become interested in the seas through their travels, skin diving, reading, or other means. It is for this diversified audience that Dr. Shepard has written this book. Throughout the book the use of technical language is minimized so that the educated layman will have no difficulty in understanding the various concepts introduced. Nor is the scientific value of the book minimized by restricting the wide use of the technical vocabulary.

The Earth Beneath the Sea is based not only on the work accomplished by the author and his associates at Scripps Institution of Oceanography but also draws extensively from the more recent work of other scientists from this country and abroad. Many interesting first-hand experiences of the author are included in the book.

The book contains eleven chapters, each with numerous subdivisions. These chapters are: (1) Waves and Currents Modify the Sea Floor, (2) Catastrophic Waves from the Sea, (3) Our Transient Beaches, (4) The Continental Shelves that Surround the Lands, (5) Origin of Continental Shelves, (6) The World's Greatest Slopes, (7) Canyons of the Sea Floor, (8) The Deep Ocean Floor, (9) Under the Ocean Bottom, (10) Coral Reefs and Their Undersea Wonderlands, and (11) Using the present Sea-Floor deposits to Interpret the Past. The book also contains a geological time scale, a list of suggested additional readings, and an index. Of particular interest to geologists concerned with sedimentation and stratigraphy is the final chapter which examines the sediments forming in particular environments today.

Throughout the book Dr. Shepard shows how the data collected by oceanographers in the past decade or past two decades have caused a revision of many popular and traditional theories concerned with the oceans, their currents, sea-floors, and the sediments of the oceans. He presents the evidence on which these theories were based and then enumerates how more recent data have caused the alteration or complete abandonment of many of these theories. Nor is the reader left with the impression that the theory or theories popular today are entirely correct; instead he will realize that other theories may be needed as more data are collected.

The book is excellent reading for anyone interested in the oceans. It is hoped that the more up-to-date information that is brought together in this book will soon spread to our geology textbooks so that beginning geology students will no longer be subjected to the dated presentations which they too often encounter.

FREDERICK W. CROPP
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ERRATA

MINERALOGY AND PETROGRAPHY OF SOME MARINE BOTTOM SEDIMENT SAMPLES OFF THE COAST OF PERU AND CHILE

E-AN ZEN, 1959, Jour. Sedimentary Petrology, v. 29, p. 513-539.

Delete the superscript [2] following the author's name at the head of the article on p. 513.

Footnote 2, on p. 513, should be on p. 531.

In table 5, p. 534, the column head reading "Heated to 350°C" should read "Heated to 450°C".

JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 30, NO. 1, P. 162
MARCH, 1960

ANNOUNCEMENTS

SOCIETY OF ECONOMIC PALEONTOLOGISTS AND MINERALOGISTS OFFICERS FOR THE YEAR, APRIL, 1960–APRIL, 1961

President: William M. Furnish, Jr.
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Co-Editors, Journal of Paleontology: Marcus L. Thompson and Charles W. Collinson

At the Business Meeting of the S.E.P.M' last year in Dallas, Texas (March, 1959)' your Council was requested to poll the entire voting membership of the Society, requesting that they indicate whether they preferred a single or double slate of candidates for S.E.P.M. offices. This Opinion Referendum was mailed to all active members on August 31, 1959. The results, as of March 1, 1960, are as follows:

167—favored a single slate of candidates
293—favored 2 candidates

The Council is, therefore, bound by the majority vote, to instruct the nominating committee to submit a ballot of 2 nominees for each of the offices of President, Vice-President, and Secretary-Treasurer.

NINTH ANNUAL CLAY CONFERENCE

The Ninth Annual Clay Conference will be held on October 6, 7, and 8, 1960, at Purdue University, Lafayette, Indiana, under the auspices of the Clay Minerals Committee of the National Academy of Sciences-National Research Council.

Two symposia of invited papers will be held on the topics of "Engineering Aspects of Physico-Chemical Properties of clays" and "Clay-Organic Complexes." In addition to these special symposia there will be general sessions of contributed papers. All those having contributions should contact Dr. J. L. White, Chairman, Ninth National Clay Conference, Agronomy Department, Purdue University, Lafayette, Indiana. The title and an abstract of approximately 250 words should be sent in by June 1.

Further information and a preliminary announcement of the Conference may be obtained by writing to Dr. White.

ANNUAL MEETING AND FIELD TRIP
PERMIAN BASIN SECTION
SOCIETY OF ECONOMIC PALEONTOLOGISTS AND MINERALOGISTS
ABILENE, TEXAS, APRIL 14, 15, & 16, 1960

The annual technical and business meetings of the Permian Basin Section, Society of Economic Paleontologists and Mineralogists, will be held in Abilene, Texas, on April 14. The annual field trip, on April 15 and 16, will be held on the Bend Arch and will be devoted to a study of the lower Permian and upper Pennsylvanian rocks of that area. Meeting and Field Trip headquarters will be at the Windsor Hotel, Abilene. Mr. Thomas Hambleton of Texaco, Inc., Box 1270, Midland, Texas, is general chairman of the field trip and Dr. John P. Brand, Texas Technological College, Lubbock, Texas, is general chairman of the annual meetings.

RAYMOND W. RALL, *Secretary*
Permian Basin Section

JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 30, NO. 1, PP. 163–186
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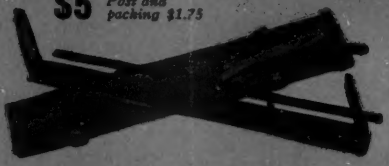

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